NAS 1.55:3115/v.2

NASA Conference Publication 3115, Vol. 2

AUG 01 [illegible]

West Antarctic Ice Sheet Initiative

Volume 2: Discipline Reviews

UNITED STATES OF AMERICA
ANTARCTIC PROGRAM

Proceedings of a workshop held at
Goddard Space Flight Center
Greenbelt, Maryland
October 16–18, 1990

nsf

NASA

NASA Conference Publication 3115, Vol. 2

West Antarctic Ice Sheet Initiative

Volume 2: Discipline Reviews

Edited by
Robert A. Bindschadler
NASA Goddard Space Flight Center
Greenbelt, Maryland

Proceedings of a workshop cosponsored by the
National Aeronautics and Space Administration,
Washington, D.C., and the National Science
Foundation, Washington, D.C., and held at
Goddard Space Flight Center
Greenbelt, Maryland
October 16–18, 1990

NASA
National Aeronautics and
Space Administration
Office of Management
Scientific and Technical
Information Program
1991

I

BLANK PAGE

PREFACE

This report contains seven discipline review papers on the state of our knowledge of West Antarctica and opinions on how that knowledge must be increased to predict the future behavior of this ice sheet. The papers were prepared to accompany lectures given at the second SeaRISE workshop held October 16-18, 1990 at NASA/Goddard Space Flight Center in Greenbelt, Maryland. The purpose of the workshop was to draft a Science and Implementation Plan of what was once called SeaRISE but is now called the West Antarctic Ice Sheet Initiative (WAIS). This plan appears as Volume I of this proceedings series.

BLANK PAGE

CONTENTS

A Review of Precipitation-Related Aspects of West Antarctic Meteorology
D.H. Bromwich, A.M. Carleton, and T.R. Parish 1

Sea-Level Response to Ice Sheet Evolution: An Ocean Perspective
S.S. Jacobs 23

Terrestrial Geophysics in the SeaRISE Project
C.R. Bentley 49

Late Wisconsin and Early Holocene Glacial History, Inner Ross Embayment, Antarctica
G.H. Denton, J.G. Bockheim, S.C. Wilson, and M. Stuiver 55

Marine Record of Late Quaternary Glacial-Interglacial Fluctuations in the Ross Sea and Evidence for Rapid, Episodic Sea Level Change Due to Marine Ice Sheet Collapse
J.B. Anderson 87

Ice Cores and SeaRISE--What We Do (and Don't) Know
R.B. Alley 111

Glacial Dynamics
I. Whillans 131

A Review of Precipitation-Related Aspects of West Antarctic Meteorology

David H. Bromwich
Byrd Polar Research Center
The Ohio State University
Columbus, Ohio 43210

Andrew M. Carleton
Department of Geography
Indiana University
Bloomington, Indiana 47405

Thomas R. Parish
Department of Atmospheric Science
University of Wyoming
Laramie, Wyoming 82071

SeaRISE Workshop Contribution
9 October 1990.

SUMMARY

An overview is presented of the factors associated with snowfall over the West Antarctic Ice Sheet. The flux of atmospheric moisture across the coast, the synoptic proceses over the South Pacific Ocean, the large-scale atmospheric controls, and numerical modeling of the West Antarctic environment are all discussed. Suggestions are made for research needed to substantially upgrade the status of knowledge in these closely-interrelated topic areas.

1. INTRODUCTION

The goal of this overview is to provide a summary of the status of knowledge concerning the primary atmospheric input to the West Antarctic ice sheet, namely snowfall. The document follows the organization of the SeaRISE document (Bindschadler, 1990). The atmospheric transport of water vapor across the coast from the Southern Ocean, whose convergence equals the average net precipitation rate over the ice sheet, is discussed in Section 2. Study of this quantity is needed because there are substantial obstacles to directly measuring snowfall over polar ice sheets (e.g., Bromwich, 1988), and this water balance approach provides a way to calculate the rate from variables which are usually readily available.

The atmospheric water balance provides a composite description of snowfall over the ice sheet, but does not resolve the individual precipitation events. Cyclonic activity causes the spatial patterns and temporal variability of the moisture fluxes, and these processes are surveyed in Section 3. Synoptic processes express the broader scale atmospheric, oceanic and ice sheet conditions. An understanding is needed of the dominant patterns of variability and how they interact in order to deduce correctly the record of past large-scale atmospheric patterns from synoptically-dominated records of individual ice cores. Section 4 addresses such aspects, focussing on El Niño Southern Oscillation (ENSO) events which are manifested on the multiannual time scale. Numerical models are primary tools in the study of the atmospheric processes that determine the precipitation and other aspects of the Antarctic climate both at present and for different boundary conditions (i.e., past and future). Global and regional models are applicable, and their status is discussed in Section 5. The final section offers some thoughts on research needed to advance the state-of-the-art.

2. MOISTURE FLUX INTO WEST ANTARCTICA

This evaluation must first consider the situation for the Antarctic continent as a whole. In general, the available estimates of poleward moisture transport from atmospheric measurements are deficient. Bromwich (1990) compared the annual moisture transport convergence values (= annual net precipitation) for Antarctica from two recent climatological atmospheric analyses with those inferable from the multiannual accumulation analysis of Giovinetto and Bentley (1985). For the continent as a whole the atmospheric-derived estimates were only one-third to one-half the terrestrial values; for 80-90°S the two approaches yielded comparable results. It was concluded that the primary causes of the large discrepancy for the continent were deficiencies in atmospheric diagnoses of the transport contributions of cyclones and surface winds near the coast. A secondary cause was the small systematic underestimate by radiosondes of the atmospheric moisture content during the cold polar night.

Table 1 (from Giovinetto et al., 1991) provides a historical overview of the estimated water vapor transports across 70°S from atmospheric and surface-based observations. The values are arranged by publication date, and only analyses drawing upon data collected during or after the International Geophysical Year (IGY) are considered. Pre-IGY evaluations were primarily conjectural because of very limited observations. There is a marked contrast between the terrestrial and atmospheric estimates. The former reveal little trend with publication year, fluctuating around a poleward transport of 6.0 kg $m^{-1}s^{-1}$. With steadily improving spatial coverage and enhanced measurement techniques, the accumulation-based estimates can be regarded as fairly reliable. The atmospheric values show a strong trend with publication year, changing from weakly equatorward to a poleward transport of 5.3 kg $m^{-1}s^{-1}$, illustrating the above weaknesses of the atmospheric analyses. Of course, the variable record lengths for the atmospheric studies complicates this comparison. It can be concluded that although the atmospheric analyses show steady improvement they are not yet suitable for continent-wide studies of the atmospheric water balance. However, for 80-90°S and perhaps a large fraction of East Antarctica (Bromwich, 1990) they can be used for this purpose.

The explanation for the above results can be sought in part in the distribution of radiosonde stations which are the primary source for the wind and moisture profiles from which moisture transports are calculated. Figure 1 (from Bromwich, 1990) describes the array of radiosonde stations available for the climatological atmospheric analyses of Peixoto and Oort (1983), and Howarth (1983) and Howarth and Rayner (1986): PO and HR respectively. The situation is rather similar today. The coverage is adequate for determination of poleward moisture transports along most of the East Antarctic coast and around the 80° latitude circle. The coast of West Antarctica is unmonitored, and the upper air station at Byrd has not operated since the early 1970s. It is probable that the former gap contributed significantly to the deficiencies of the atmospheric analyses in Table 1. The temporal trend of increasing atmospheric poleward moisture transports in Table 1 corresponds to the degree of monitoring of atmospheric processes over the ocean areas surrounding the continent. During the IGY-winter the Southern Ocean was almost completely devoid of observations. The Peixoto and Oort (1983) analysis relied on radiosonde data, but the Howarth (1983) and Howarth and Rayner (1986) diagnosis profited from the incorporation of oceanic satellite observations (from imagery and soundings). During 1979, there was a comprehensive drifting buoy program in the Southern Ocean (Hollingsworth, 1989), and this data source appears to have been one key to the markedly improved analysis by Masuda (1990).

Studies focussing specifically on moisture transports into West Antarctica were conducted by Lettau (1969) and Rubin and Giovinetto (1962), and primarily drew upon IGY-era data. The presence of upper air stations at Byrd and Little America V (northeastern edge of the Ross Ice Shelf) was a critical aspect for both studies. From a continental atmospheric mass budget and assumed moisture values, Lettau (1969) inferred that 40% of the vapor transported into Antarctica enters in this sector with the time-averaged and storm transports contributing equally. Notice that no direct moisture budget was attempted because of skepticism about the quality of the radiosonde humidity measurements at low air temperatures; the present authors do not share this viewpoint but do recognize the need for caution in using such data. The importance of cyclonic events for West Antarctic snowfall was underscored by the work of Vickers (1966), and implies that interannual variability may be high. Rubin and Giovinetto (1962) showed that the broad spatial patterns of snow accumulation were consistent with the moisture flux directions and humidity characteristics at radiosonde stations surrounding the ice sheet; once again no moisture budget analysis was attempted.

Although recent atmospheric analyses need to be checked it appears that to conduct meaningful diagnostic analyses of the atmospheric moisture budget over West Antarctica the data base must be enhanced. The most effective action would be to improve the oceanic drifting buoy program so that coverage approaches that during 1979 (see Hollingsworth, 1989). Radiosonde programs (or equivalent) should be implemented at any manned station set-up by the United States in West Antarctica to replace Siple. Finally, satellite remote sensing offers much potential, and vigorous efforts are underway to solve the difficulties associated with soundings over sea-ice and ice-sheet surfaces (e.g., Claud et al., 1988).

3. SOUTH PACIFIC SYNOPTIC ACTIVITY

Much of our understanding of synoptic processes over the extensive middle and high latitude oceans of the Southern Hemisphere is based on only about 30 years of widely-spaced station sea-level-pressure (SLP) observations. These comprise principally the following climatologies:

(a) The 18-month period of the IGY of 1957-58 (e.g., Astapenko, 1964; Taljaard, 1967).

(b) The South African hemispheric analyses that continued into the mid-1960s (Taljaard et al., 1969; Newton, 1972).

(c) The Australian Numerical Meteorology Research Center (ANMRC) hemispheric analyses that began in the early 1970s (Streten, 1980a; LeMarshall and Kelly, 1981; LeMarshall et al., 1985).

(d) The European Center for Medium-Range Weather Forecasting (ECMWF) global-scale analyses dating from the First GARP Global Experiment (FGGE) of 1979.

With specific regard to the South Pacific, the coverage was particularly deficient in (b), and only improved from the early 1970s; first with the regular incorporation of satellite information (c), and then with the inclusion of ocean drifting buoys and automatic weather station (AWS) data for the Antarctic (d). These SLP analyses reveal the following general features of the synoptic atmospheric circulation.

(1) On a monthly and seasonal basis, a low amplitude long wave trough is located through the South Pacific in association with a relatively zonal circulation (Trenberth, 1979-- his Fig. 1; Streten and Troup, 1973). Pressures decrease into the circumpolar trough, which is located between about 60° to 74°S, depending on longitude considered and also month/season (Streten, 1980a-- his Figs. 12, 13). For West Antarctica, the circumpolar trough is generally located between about 68-72°S (Fig. 2).

(2) On shorter (daily) time scales, the circulation may be considerably more meridional, with high pressure ridges interrupting the circumpolar trough. Preferred longitudes for this activity in the West Antarctic sector are between about 80-70°W, and concentrated in the autumn and winter (Streten, 1980a-- his Fig. 15). However, this activity is considerably reduced compared with that of cyclones, which peaks between about 170-130°W in this region (Fig. 2).

(3) The circumpolar trough is the summation of individual frontal cyclones moving in from middle latitudes, which dissipate and stagnate preferentially in the major Antarctic embayments. In the West Antarctic these are the eastern Ross Sea, the Amundsen/Bellinghausen seas, and the Weddell Sea (Fig. 2).

(4) The circumpolar trough undergoes a dominant semi-annual oscillation in position and intensity. It is closest to Antarctica and deeper in the equinoctial months (January, June) compared with the solsticial months (March, September) (Streten, 1980a-- his Fig. 12). Similarly, the strength of the zonal index (westerlies) for the latitude zone 40-60°S is highest in February/March and September/October (Streten, 1980a-- his Fig. 10). In this regard, the sector from 180° eastward to 90°W is no exception (Streten, 1980a-- his Fig. 9). The semi-annual oscillation is somewhat reflected in the seasonal distribution of precipitation around West Antarctica (van Loon, 1972; Bromwich, 1988). The semi-annual oscillation involves changes in the hemispheric waves, particularly over the Indian and South Pacific oceans and in the colder part of the year (van Loon and Rogers, 1984). These changes are, in large part, responsible for the "coreless" winter observed at many Antarctic stations, particularly in the Ross Sea sector. Over middle latitudes, the changes in the waves associated with the semi-annual oscillation are manifest in marked changes in the preferred longitudes of polar lows between June and September (Carleton and Carpenter, 1990).

(5) On an interannual basis the circumpolar trough undergoes a marked variation in mean latitude for the Ross, Amundsen and Bellingshausen sea sectors. This exceeds 8° of latitude for the early autumn, early winter and late spring, but is much smaller (less than 4°) in the summer and late winter/spring, at least for the period 1972-77 (Streten, 1980a).

(6) Anomalies of SLP for the Pacific are often out-of-phase between low/high and middle latitudes, and between low and high latitudes (Mo and White, 1985). These teleconnections are associated with the Southern Oscillation, particularly in summer. They are associated with strengthening and weakening of the westerlies in alternating latitude belts (Rogers and van Loon, 1982), and are such that stronger (weaker) trades tend to have associated stronger (weaker) westerlies north (south) of about 45°S (Trenberth, 1981). In winter, SLP anomalies are often out-of-phase between the Australasian and Antarctic

Peninsula/South American regions. These correspond with the Trans-Polar Index (TPI) of Pittock (1984), which is the SLP anomaly difference between Hobart and Stanley. Variations in TPI correlate with variations in Scotia Sea pack ice conditions (Rogers and van Loon, 1982).

(7) On decadal time scales, there are variations apparent in the frequencies of cyclones and their mean longitudes of occurrence for the West Antarctic. Data for the IGY and subsequent years (Taljaard et al., 1969) show a weaker mean low in the vicinity of the Ross and Weddell seas compared with the later climatologies (Streten, 1980a), particularly in the 1978-82 period (LeMarshall et al., 1985). Thus, there have also been variations within the time period of the ANMRC 10-year climatology (LeMarshall et al., 1985). These point up the highly variable nature of the Southern Hemisphere circulation on a range of time scales, even given the effects of increased data and the changes in analysis procedure (LeMarshall and Kelly, 1981).

4. ENSO RELATIONSHIPS WITH THE ANTARCTIC

The relatively short period of conventional meteorological data in the Antarctic does not facilitate a full description of ENSO teleconnections to Antarctica. However, there are tantalizing hints of the teleconnection that require further exploration as the data record lengthens, and as numerical models become more sophisticated. In particular, the analysis of ice cores should shed considerable light, assuming that the regional expression of ENSO and correlations among its descriptor variables remain relatively stable from event to event (either positive or negative phase). There are indications that this has not been the case, even for the tropical centers of action (Elliot and Angell, 1988). The following are some important observations for the Antarctic.

(1) An analysis of the time evolution of global-scale SLP anomalies (Krishnamurti et al., 1986) shows strong variance on ENSO time scales over Antarctic latitudes. In addition, a northward propagation of zonally-averaged anomalies from the south polar to north polar latitudes took place in the period 1961-71, but reversed after that time. Additional support for these longer-term Antarctic influences on global climate is given in Fletcher et al.'s (1982) study of historical sea-surface temperature and surface-wind observations over the Southern Ocean south of Australia (40-50°S).

(2) A hint of a quite strong higher latitude manifestation of ENSO is apparent in hemispheric composites of SLP anomalies associated with the peak years of warm events compared with the years immediately prior (van Loon and Shea, 1985). These suggest the existence of quite strong pressure reversals over the Weddell Sea, south Indian Ocean, and possibly also over the southeast Pacific south of 50°S (their Figs. 1a, b). They were confirmed, at least for the Weddell Sea sector, by Carleton (1988), who also showed the existence of substantial associated anomalies in sea ice concentration for that region.

(3) Regional-scale anomalies in the sea ice extent and concentration of the Antarctic embayments are strongly associated with changes in temperature, pressure and winds -- hence, the role of cyclonic activity noted earlier.

Accordingly, changes in sea ice conditions (extent, concentration) are an indicator of climate anomalies. Chiu (1983) finds a significant association between the SOI (Southern Oscillation Index) in March and April and the Antarctic sea ice area in the following July-December period, and also an apparent lag of the SOIs with sea ice area in the latter part of the year. The study was extended by Carleton (1989), who considered regional-scale sea ice changes associated with ENSO and the hemispheric long waves. There is also an apparent ENSO signature in the patterns of subsynoptic-scale cyclogenesis ("polar lows") over the South Pacific and southern Indian Oceans, apparently in response to the large-scale changes in the long waves and associated outbreaks of colder air toward lower latitudes (Carleton and Carpenter, 1990).

(4) Using the period of record for the South Pole and AWS data in the Ross Sea area for the 1982-83 ENSO event, Savage et al. (1988) identify a lagged effect of ENSO in Antarctic temperatures and surface winds (Fig. 3a, b). Very low temperatures at South Pole apparently resulted in enhanced cold air drainage and higher katabatic wind speeds reported near the coast at that time.

(5) Available general circulation modeling studies (e.g., Mitchell and Hills, 1986; Simmonds and Dix, 1986) suggest quite strong and, for the most part, significant tropical pressure and height responses to prescribed anomalies of the Antarctic wintertime sea ice extent. These appear to have an ENSO signature. Changes in cyclonic activity occur in the circumpolar trough; however, the magnitude and even sign of the SLP departures appears to be model dependent (cf., Mitchell and Hills, 1986; Simmonds and Dix, 1986).

It is apparent from longer-term indices of ENSO that the 'poles' of the Southern Oscillation and the frequency of high magnitude events have changed during the past 100 years or so. An index of the Oscillation back to 1600 A.D. and derived from analysis of tree rings, suggests even longer-time scale changes in the frequency of low-index (warm) events (Lough and Fritts, 1985). These appear to have been less frequent in the nineteenth century compared with the eighteenth and twentieth centuries. Also, the warm events of 1792-93 and 1815-16 may have been of comparable magnitude to the major 1982-83 event. The analysis of Antarctic ice core data could be used to extend the ENSO record back even further. However, this assumes (a) the existence through time of a stable and coherent teleconnection to the Antarctic circulation, precipitation and temperature regimes; and (b) the identification of ice-core drilling sites within such center(s) of action. From the foregoing discussion, the Pacific sector of the West Antarctic appears to be such a place; however, the dominance of the annual snowfall regime by relatively few cyclonic events is problematic. A candidate mechanism for the tropical-Antarctic teleconnection is the South Pacific Cloud Band (SPCB). This feature is an important avenue for the transports of energy and moisture, and also has a strong ENSO signature (Streten, 1975; Trenberth, 1976; Meehl, 1988). The possible impact of such low-frequency changes in the SPCB for the accumulation on the West Antarctic ice sheet is unknown at this time.

5. NUMERICAL MODELING OF THE WEST ANTARCTIC ENVIRONMENT

Studies using atmospheric general circulation models (GCM) and high resolution regional models are needed to improve our understanding of the planetary, synoptic and mesoscale atmospheric processes over Antarctica. Incorporation into GCMs of knowledge gained from diagnostic studies such as those outlined above will make it possible to commence a realistic assessment of the interactions between Antarctica and the rest of the planet. Complete exploration of this topic will require coupling of GCMs to realistic ocean and sea-ice models.

An early assessment of the ability of GCMs to reproduce the present Antarctic climate (Schlesinger, 1984) revealed large systematic errors in temperature, pressure, precipitation and cyclonic simulations. Model deficiencies of this magnitude preclude their use to understand the present climate or to describe how it might have changed in the past or might do so in the future. This situation persists for some models, as indicated by the NCAR Community Climate Model being unable to simulate major Antarctic environmental changes within the last 3 million years (Elliot et al., 1991). Simmonds (1990) demonstrates that the veracity of many models has continued to improve through the end of the 1980s, and presents convincing simulation fields of sea-level pressure, strength of the surface temperature inversion over the ice sheets during winter, and annual precipitation, to illustrate this contention. As an example, Figure 4 taken from Mitchell and Senior (1989) shows a qualitatively reasonable simulation of the Antarctic surface winds by the British Meteorological Office GCM. Simmonds points out that it is difficulty to verify models because real climatic variables are not known to a high degree of confidence. This argues for a vigorous effort to fill in the large gaps in the observational network, particularly over the South Pacific Ocean, the coast of West Antarctica, and the interior of the ice sheet.

Numerical simulation of high latitude atmospheric phenomena using sophisticated, high resolution models is an extremely promising avenue of future research. Mesoscale (10-300 km) processes such as katabatic wind circulations (Parish, 1984; Parish and Bromwich, 1986, 1987, 1991) and polar lows (Bromwich, 1989; 1991) are frequently observed in Antarctic regions and may play a significant role in the heat and momentum exchanges between polar and tropical latitudes. Such processes are at best only crudely parameterized within the context of GCMs (see, for example, Mitchell and Senior, 1989); it is extremely important that effects of such mesoscale processes on larger-scale circulations and transports be quantified before definitive GCM studies can be undertaken.

To date, a large gap exists in modeling applications of regional-scale Antarctic processes. Only specialized applications involving mesoscale models have been attempted. As an example, mesoscale models have been used to understand the temporal and spatial evolution of katabatic winds. Parish and Waight (1987) have used a two-dimensional numerical model to simulate the development of katabatic drainage. Emphasis was placed on the evolution of radiative and sensible heat transports in the lower atmosphere on the forcing of the katabatic flow. Parish and Wendler (1990) have used a hydrostatic, three-dimensional model to depict the forcing of the anomalous katabatic wind regime near Adelie Land, Antarctica. A similar katabatic wind study has been completed by van Meurs and Allison (1989) for a section of East Antarctica along 112°E. Parish and Bromwich (1991) have used a numerical model to portray the katabatic wind pattern over the Antarctic continent. The irregularity of the katabatic drainage pattern into "confluence zones" modulates the intensity of coastal katabatic winds and is thought to influence the cyclogenetic potential of the near-coastal periphery (Bromwich, 1991). Most of these model simulations encompass relatively short time periods on the order of 12 hours to a few days. Longer time-scale studies will be necessary to integrate these regional simulations with GCM results. For example, Egger (1985) and James (1989) have noted that the katabatic wind circulation is tied directly to the large-scale circumpolar vortex about the Antarctic continent. Using simplified models to represent the radiative cooling of the sloping terrain, both authors showed that the shallow katabatic winds were responsible for the development of an upper tropospheric vortex on a time-scale of approximately several days to a week or more. They conclude that the katabatic wind circulations must act in unison with propagating cyclones in the evolution of the upper level vortex. Intensification of the upper level vortex pattern tends to suppress further katabatic wind development; cyclonic intrusion is thought to disrupt the circumpolar vortex which then allows the katabatic wind circulation to become reestablished. Thus, a significant scale-interaction is suggested.

A critical component of future numerical studies will be the incorporation of extratropical cyclones and the resulting moisture and heat fluxes into mesoscale models. As shown by Bromwich (1988), a significant moisture flux onto the Antarctic continent is the result of cyclonic activity. Inclusion of such processes within detailed regional models is prerequisite to detailed GCM studies of Antarctic ice sheet variability and climate change.

6. CONCLUSIONS

Implicit to the preceding discussion is that West Antarctic meteorology must be viewed within the context of the atmospheric dynamics of high southern latitudes, and that the broadscale ice sheet configurations in both the eastern and western hemispheres coupled with the general atmospheric circulation results in the atmospheric processes governing snowfall over West Antarctica. Central to the advancement of basic knowledge is the need to eliminate the data void over the South Pacific Ocean and West Antarctica. Strategies to address this problem include deployments of AWS on the ice sheet and on offshore islands, and of free-floating buoys in the ocean areas. Radiosonde programs should be instigated at at all manned stations established in the future. Polar orbiting satellite data can help fill the data gap, and vigorous exploitation of the satellite imagery and soundings routinely collected at McMurdo and Palmer stations is needed. These two sites can provide complete high resolution coverage of West Antarctica several times a day.

For the moisture fluxes the first requirement is to establish an adequate observational system so that the temporal and spatial modes can be characterized and related to the large-scale atmospheric circulation, synoptic processes and oceanic boundary conditions (e.g., sea ice and sea-surface temperatures). A deeper quantitative understanding of these interactions is critical because it will allow accumulation variations, which are difficult to measure everywhere on the ice sheet, to be inferred from more viable measurements of the factors that determine the accumulation pattern.

Very little is known about cyclonic processes over West Antarctica. There is a need to characterize the formation, tracks and dissipation of cyclones in this area together with the associated variability; both synoptic and mesoscale cyclones may be important and may interact in significant ways. Of particular importance is an examination of the frequency of and mechanisms by which cyclones penetrate deep into the interior of the ice sheet. A 10- to 15-year daily synoptic data base established from all available observations is required for such studies, and is becoming feasible.

The connections between the large-scale and synoptic-scale processes must be understood on time scales ranging from weekly and seasonal through interannual to decadal. As above, a 10- to 15-year combination of satellite and operational analyses is needed to describe the variability of the large-scale processes and the interactions among the component parts.

Interactive evaluations of numerical modeling/theoretical studies and observational analyses are required to extract the maximum understanding and realism from all approaches. Understanding the important processes and their sensitivity to orographic, sea-ice and sea-surface temperature characteristics as well as the CO_2 content of the atmosphere is vital for the study and prediction of the West Antarctic ice sheet in particular, and of sea-level changes in general.

ACKNOWLEDGMENTS. Preparation of this overview was supported by NSF grants DPP-8916921 (DHB), DPP-8816912 (AMC) and DPP-8916998 (TRP).

REFERENCES

Astapenko, P.D., 1964. Atmospheric processes in the high latitudes of the Southern Hemisphere. Section II of *IGY Program (Meteorology)*, No. 3. Israel Program for Scientific Translations, 286 pp.

Baumgartner, A. and E. Reichel, 1975. *The World Water Balance*. Elsevier, New York.

Bindschadler, R.A. (ed.), 1990. *SeaRISE: a multidisciplinary research initiative to predict rapid changes in global sea level caused by collapse of marine ice sheets*. NASA Conference Publication 3075. National Aeronautics and Space Administration, Scientific and Technical Information Division, 47 pp.

Bromwich, D.H., 1988. Snowfall in high southern latitudes. *Rev. Geophys.*, 26, 149-168.

Bromwich, D.H., 1989. Subsynoptic-scale cyclone developments in the Ross Sea sector of the Antarctic. In: Twitchell, P.F., E.A. Rasmussen and K.L. Davidson (eds.), *Polar and Arctic Lows*. A Deepak Publishing, Hampton, Virginia, 331-345.

Bromwich, D.H., 1990. Estimates of Antarctic precipitation. *Nature*, 343, 627-629.

Bromwich, D.H., 1991. Mesoscale cyclogenesis over the southwestern Ross Sea linked to strong katabatic winds. *Mon. Wea. Rev., in press.*

Carleton, A.M., 1988. Sea-ice atmosphere signal of the Southern Oscillation in the Weddell Sea, Antarctica. *J. Climate*, 1, 379-388.

Carleton, A.M., 1989. Antarctic sea-ice relationships with indices of the atmospheric circulation of the Southern Hemisphere. *Climate Dynamics*, 3, 207-220.

Carleton, A.M. and D.A. Carpenter, 1990. Satellite climatology of "polar lows" and broadscale climatic associations for the Southern Hemisphere. *Intl. J. Climatol.*, 10, 217-246.

Chiu, L.S., 1983. Antarctic sea ice variations 1973-1980. In: Street-Perrott, A., M. Beran and R. Ratcliffe (eds.), *Variations in the Global Water Budget*. Reidel, Dordrecht, 301-311.

Claud, C., A. Chedin, N.A. Scott, J.C. Gascard and G.J. Prangsma, 1988. Retrieval of mesoscale meteorological parameters from NOAA-9 over Greenland and Barents seas for three consecutive days of June 1986. In: Preprints volume, *Second Conference on Polar Meteorology and Oceanography*, Madison, Wisconsin: March 29-31, 1988. Amer. Meteor. Soc., Boston, Mass., 122-125.

Egger, J., 1985. Slope winds and the axisymmetric circulation over Antarctica. *J. Atmos. Sci.*, 42, 1859-1867.

Elliot, D.H., D.H. Bromwich, D.H. Harwood and P.-N. Webb, 1991. The Antarctic glacial geologic record and GCM modeling: a test. *Proceedings of the International Conference on the Role of the Polar Regions in Global Change*, Fairbanks, Alaska: June 11-15, 1990, *in press*.

Elliot, W.P. and J.K. Angell, 1988. Evidence for changes in Southern Oscillation relationships during the last 100 years. *J. Climate*, 1, 729-737.

Fletcher, J.O., U. Radok and R. Slutz, 1982. Climatic signals of the Antarctic Ocean. *J. Geophys. Res.*, 87, 4269-4276.

Giovinetto, M.B. and C.R. Bentley, 1985. Surface balance in ice drainage systems of Antarctica. *Antarct. J. U.S.*, 20(4), 6-13.

Giovinetto, M.B., D.H. Bromwich and G. Wendler, 1991. Atmospheric net water vapor and latent heat transport across 70°S. *J. Geophys. Res., in review.*

Hollingsworth, A., 1989. The Global Weather Experiment - 10 years on. *Weather*, 44, 278-285.

Howarth, D.A., 1983. Seasonal variations in the vertically integrated water vapor transport fields over the southern hemisphere. *Mon. Wea. Rev.*, 111, 1259-1272.

Howarth, D.A. and J.N. Rayner, 1986. Estimates of sources and sinks of atmospheric moisture in the Southern Hemisphere. In: Preprints volume, *Second International Conference on Southern Hemisphere Meteorology*, Wellington, N.Z.: Dec. 1-5, 1986. Amer. Meteor. Soc., Boston, Mass., 163-166.

James, I.N., 1989. The Antarctic drainage flow: Implications for hemispheric flow on the Southern Hemisphere. *Antarct. Sci.*, 1, 279-290.

Krishnamurti, T.N., S-H. Chu and W. Iglesias, 1986. On the sea level pressure of the Southern Oscillation. *Arch. Meteor. Geophys. Bioklim.*, 34A, 385-425.

LeMarshall, J.F. and G.A.M. Kelly, 1981. A January and July climatology of the southern hemisphere based on daily numerical analyses, 1973-1977. *Austral. Meteor. Mag.* , 29, 115-123.

LeMarshall, J.F., G.A.M. Kelly and D.J. Karoly, 1985. An atmospheric climatology of the southern hemisphere based on ten years of daily numerical analyses (1972-82). I. Overview. *Austral. Meteor. Mag.*, 33, 65-85.

Lettau, B., 1969. The transport of moisture into the Antarctic interior. *Tellus*, 21(3), 331-340.

Lough, J.M. and H.C. Fritts, 1985. The Southern Oscillation and tree rings: 1600-1961. *J. Climate Appl. Meteor.*, 24, 952-966.

Masuda, K., 1990. Atmospheric heat and water budgets of polar regions: analysis of FGGE data. *Proc. NIPR Symp. Polar Meteor. Glaciol.*, 3, 79-88.

Meehl, G.A., 1988. Tropical-midlatitude interactions in the Indian and Pacific sectors of the Southern Hemisphere. *Mon. Wea. Rev.*, 116, 472-484.

Mitchell, J.F.B. and T.S. Hills, 1986. Sea-ice and the antarctic winter circulation: a numerical experiment. *Quart. J. Royal Meteor. Soc.*, 112, 953-969.

Mitchell, J.F.B. and C.A. Senior, 1989. The antarctic winter; simulations with climatological and reduced sea-ice extents. *Quart. J. Roy. Meteor. Soc.*, 115, 225-246.

Mo, K.C. and G.H. White, 1985. Teleconnections in the Southern Hemisphere. *Mon. Wea. Rev.*, 113, 22-37.

Newton, C.W. (ed.), 1972. *Meteorology of the Southern Hemisphere*. Meteor. Monograph. Vol. 13, No. 35. Amer. Meteor. Soc., Boston, Mass., 263 pp.

Parish, T.R., 1984. A numerical study of strong katabatic winds over Antarctica. *Mon. Wea. Rev.*, 112, 545-554.

Parish, T.R. and D.H. Bromwich, 1986. The inversion wind pattern over West Antarctica. *Mon. Wea. Rev.*, 114, 849-860.

Parish, T.R. and D.H. Bromwich, 1987. The surface windfield over the Antarctic ice sheets. *Nature*, **328**, 51-54.

Parish, T.R. and D.H. Bromwich, 1991. Continental-scale simulation of the Antarctic katabatic wind regime. *J. Climate*, in press.

Parish, T.R. and K.T. Waight, 1987. The forcing of Antarctic katabatic winds. *Mon. Wea. Rev.*, 115, 2214-2226.

Parish, T.R. and G. Wendler, 1990. The katabatic wind regime at Adelie Land, Antarctica. *Intl. J. Climatol., in press.*

Peixoto, J.P. and A.H. Oort, 1983. The atmospheric branch of the hydrological cycle and climate. In: Street-Perrott, A., M. Beran and R. Ratcliffe (eds.), *Variations in the Global Water Budget*. D. Reidel, Dordrecht, 5-65.

Pittock, A.B., 1984. On the reality, stability and usefulness of Southern Hemisphere teleconnections. *Austral. Meteor. Mag.*, 32, 75-82.

Rogers, J.C. and H. van Loon, 1982. Spatial variability of sea level pressure and 500 mb height anomalies over the Southern Hemisphere. *Mon. Wea. Rev.*, 110, 1375-1392.

Rubin, M.J., 1962. Atmospheric advection and the Antarctic mass and heat budget. In: Wexler, H., M.J. Rubin and J.E. Caskey, Jr. (eds.), *Antarctic Research, Geophys. Monogr. Ser.*, vol. 7. Amer. Geophys. Union, Washington, D.C., 149-159.

Rubin, M.J. and M.B. Giovinetto, 1962. Snow accumulation in central West Antarctica as related to atmospheric and topographic factors. *J. Geophys. Res.*, 67, 5163-5170.

Savage, M.L., C.R. Stearns and G.A. Weidner, 1988. The Southern Oscillation signal in Antarctica. In: Preprints volume, *Second Conference on Polar Meteorology and Oceanography*, Madison, Wisconsin: March 29-31, 1988. Amer. Meteor. Soc., Boston, Mass., 141-144.

Schlesinger, M.E., 1984. Atmospheric general circulation model simulations of the modern Antarctic climate. In: *Environments of West Antarctica: Potential CO_2-Induced Changes*. National Academy Press, Washington, D.C., 155-196.

Simmonds, I., 1990. Improvements in general circulation model performance in simulating Antarctic climate. *Antarct. Sci.*, 2, 287-300.

Simmonds, I. and M. Dix, 1986. The circulation changes induced by the removal of Antarctic sea ice in a July general circulation model. In: Preprints volume, *Second International Conference on Southern Hemisphere Meteorology*, Wellington, N.Z.: Dec. 1-5, 1986. Amer. Meteor. Soc., Boston, 107-110.

Starr, V.P., J.P. Peixoto and R.G. McKean, 1969. Pole-to-pole moisture conditions for the IGY. *Pure and Applied Geophysics*, 75, 300-331.

Streten, N.A., 1975. Satellite-derived inferences to some characteristic of the South Pacific atmospheric circulation associated with the Niño event of 1972-73. *Mon. Wea. Rev.*, 103, 989-995.

Streten, N.A., 1980a. Some synoptic indices of the Southern Hemisphere near sea level circulation 1972-77. *Mon. Wea. Rev.*, 108, 18-36.

Streten, N.A., 1980b. Antarctic meteorology: the Australian contribution past, present and future. *Austral. Meteor. Mag.*, 28, 105-140.

Streten, N.A. and A.J. Troup, 1973. A synoptic climatology of satellite-observed cloud vortices over the Southern Hemisphere. *Quart. J. Royal Meteor. Soc.*, 99, 56-72.

Taljaard, J.J., 1967. Development, distribution and movement of cyclones and anticyclones in the Southern Hemisphere during the IGY. *J. Appl. Meteor.*, 6, 973-987.

Taljaard, J.J., H. van Loon, H.L. Crutcher, and R.L. Jenne, 1969. *Climate of the Upper Air: Part 1. Southern Hemisphere*. Vol. 1, NAVAIR 50-1C-55.

Trenberth, K.E., 1976. Spatial and temporal variations of the Southern Oscillation. *Quart. J. Royal Meteor. Soc.*, 102, 639-653.

Trenberth, K.E., 1979. Interannual variability of the 500 mb zonal mean flow in the Southern Hemisphere. *Mon. Wea. Rev.*, 107, 1515-1524.

Trenberth, K.E., 1981. Interannual variability of the Southern Hemisphere 500 mb flow: regional characteristics. *Mon. Wea. Rev.*, 109, 127-136.

van Loon, H., 1972. Cloudiness and precipitation in the Southern Hemisphere. In: Newton, C.W. (ed.) *Meteorology of the Southern Hemisphere*, Meteor. Monog. Vol. 13, No. 35. Amer. Meteor. Soc., Boston, Mass., 101-111.

van Loon, H. and J.C. Rogers, 1984. Interannual variations in the half-yearly cycle of pressure gradients and zonal wind at sea level on the Southern Hemisphere. *Tellus*, 36A, 76-86.

van Loon, H. and D.J. Shea, 1985. The Southern Oscillation Part IV. The precursors south of 15°S to the extremes of the Oscillation. *Mon. Wea. Rev.*, 113, 2063-2074.

van Meurs, B. and I. Allison, 1989. The structure of katabatic flow along 112•E as inferred from model and automatic weather station results. In: Extended abstracts, *Conference and Workshop on Antarctic Weather and Climate*: July 5-7, 1989. Flinders Institute for Atmospheric and Marine Sciences, Flinders University of South Australia, 148-150.

Vickers, W.W., 1966. A study of ice accumulation and tropospheric circulation in West Antarctica. In: Rubin, M.J. (ed.), *Studies in Antarctic Meteorology*, Antarctic Research Series, Vol. 9. Amer. Geophys. Union, Washington, D.C., 135-176.

Table 1: Comparison of annual meridional water vapor transport estimates (kg m^{-1} s^{-1}) across 70°S since the International Geophysical Year (1957-1958). Negative values are directed toward the South Pole.

ATMOSPHERIC		**MULTIANNUAL SURFACE-BASED**	
+0.73	Starr et al. (1969) for 1958	-5.6 to -7.2	Rubin (1962) from accumulation data*
-3.0 (PO)	Peixoto and Oort (1983) for 1963-1973	-5.4	Baumgartner and Reichel (1975)
-3.7 (HR)	Howarth (1983) and Howarth and Rayner (1986) for 1973-1978 and 1980-1984	-5.8 to -6.0	Bromwich (1990)*
-5.3	Masuda (1990) for 1979	-6.3 ±1.1	Giovinetto et al. (1991)

*Transport across the coast of the Antarctic continent converted to transport across 70°S using ratio of the two values given by Baumgartner and Reichel (1975).

Fig. 1. Antarctic radiosonde stations whose data primarily determined the high latitude atmospheric moisture fluxes obtained by PO and HR: filled circles denote sites contributing to both studies and filled squares to only one. Notice the measurement gap along the West Antarctic coast (to the left); no upper air observations have been collected at Byrd Station (80°S) since the early 1970s. Thin continuous lines are elevation contours in meters, starting at 2000 m. Adapted from Bromwich (1990).

Fig. 2. Main schematic features of the surface circulation; inner hatched area denotes the region of the sub-Antarctic trough within which the mean seasonal axis varied between 1972 and 1977; black area shows the region of the quasi-high pressure over the interior within which monthly mean relative pressure maxima were located between 1972 and 1977; hatched histograms along 60°S represent the relative frequency of monthly mean low pressure centers in the Antarctic trough between 1972 and 1977. Adapted from Streten (1980b).

Fig. 3(a). Distribution of annual mean temperatures at Amundsen-Scott South Pole Station, Antarctica, 1957-1986. (b). Normalized time series of the annual values of the Southern Oscillation Index (SOI: Tahiti minus Darwin sea level pressure) and of the annual mean temperatures from the following year (lag +1) at the Amundsen-Scott South Pole Station. Stippling shows the seven ENSO episodes that have occurred since 1955. Adapted from Savage et al. (1988).

Fig. 4. Low level winds: (a) Model winds at $\sigma = 0.987$, July to September. The wind speed is proportional to the length of the arrows, and the scale is given in the bottom left-hand corner. (b) Time-averaged near-surface wintertime streamlines of cold air drainage over Antarctica (after Parish and Bromwich, 1987). The thin lines are ice sheet elevation contours at intervals of 100 m. Adapted from Mitchell and Senior (1989).

BLANK
PAGE

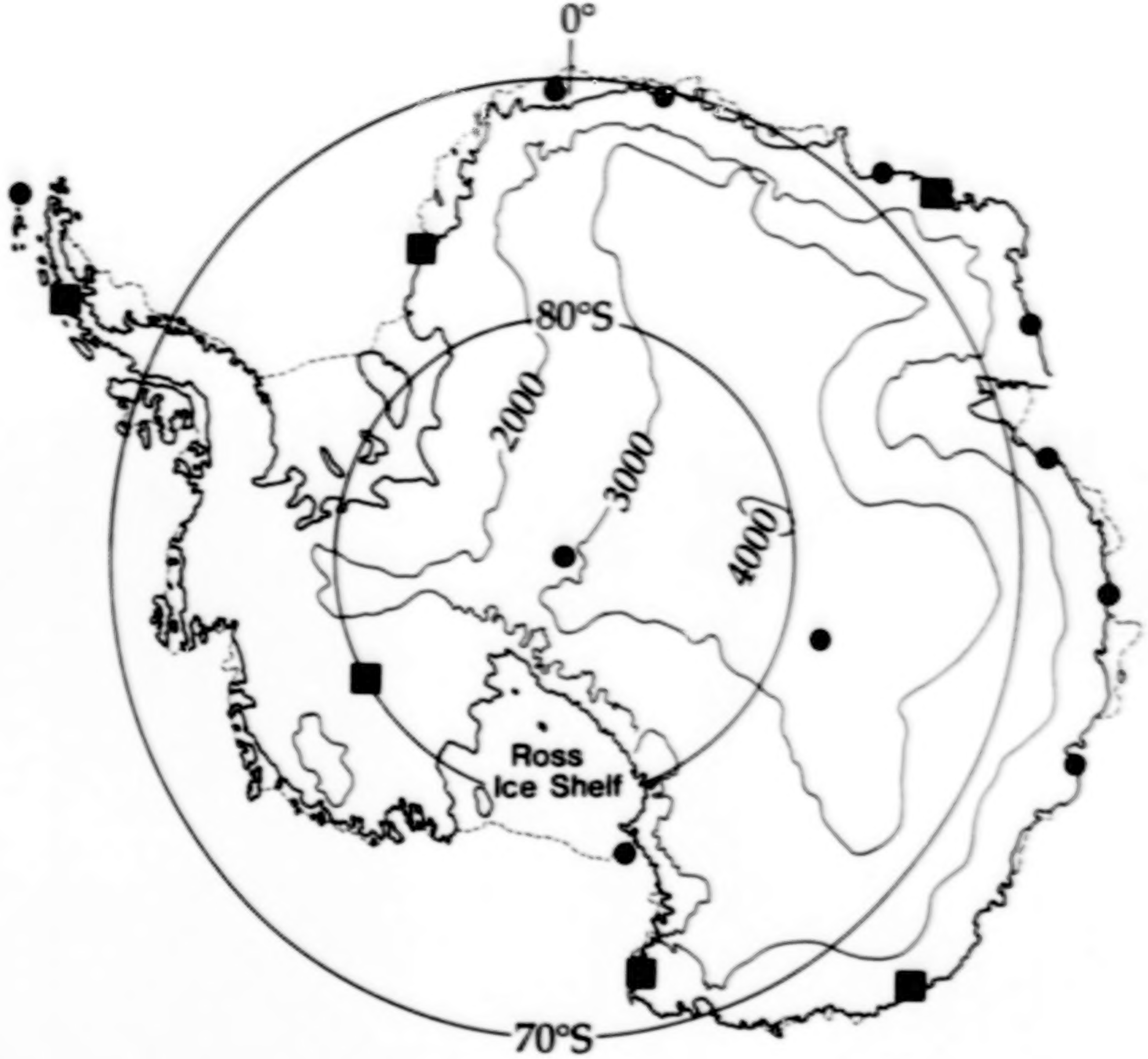


Figure 1.

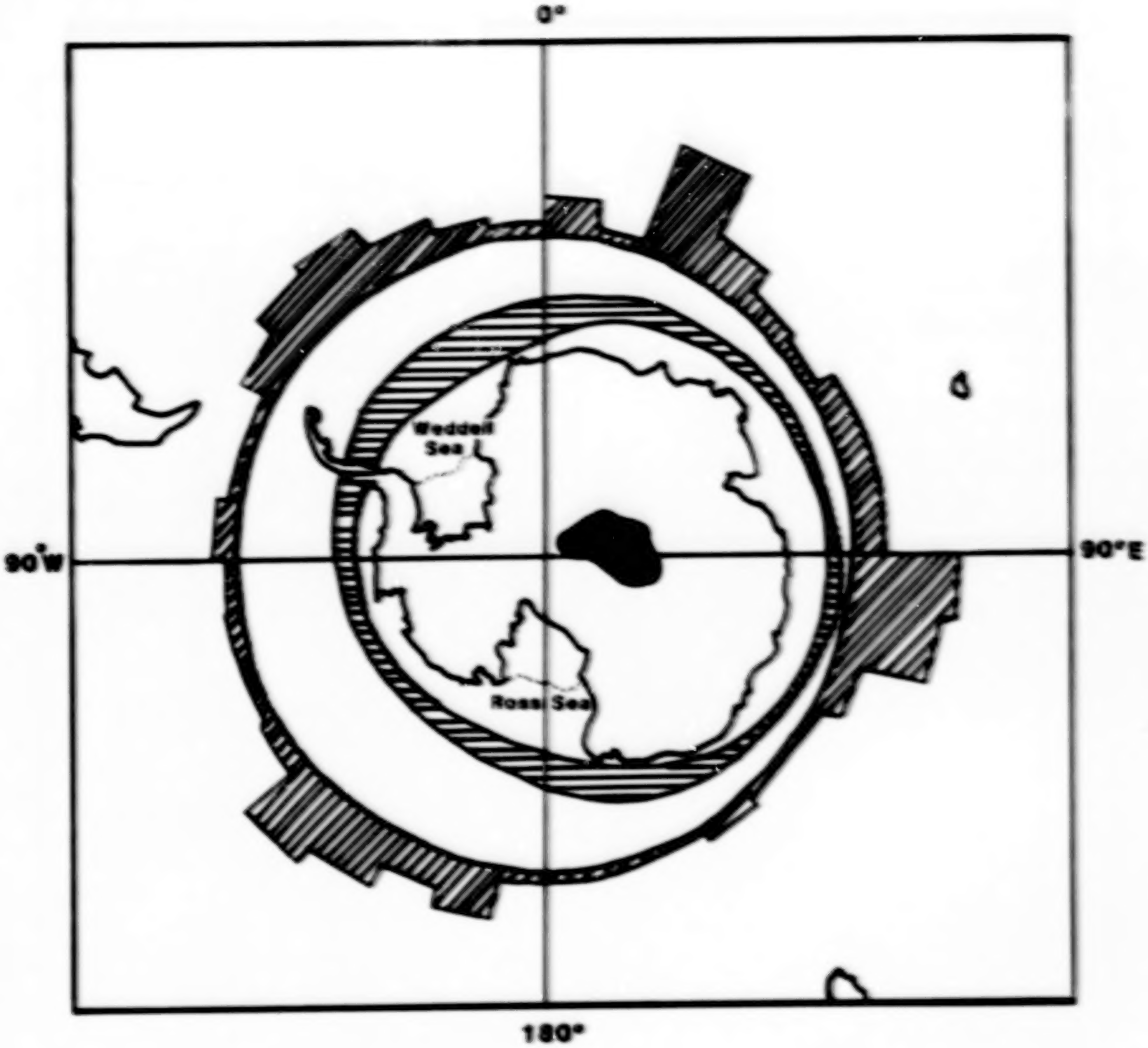


Figure 2.

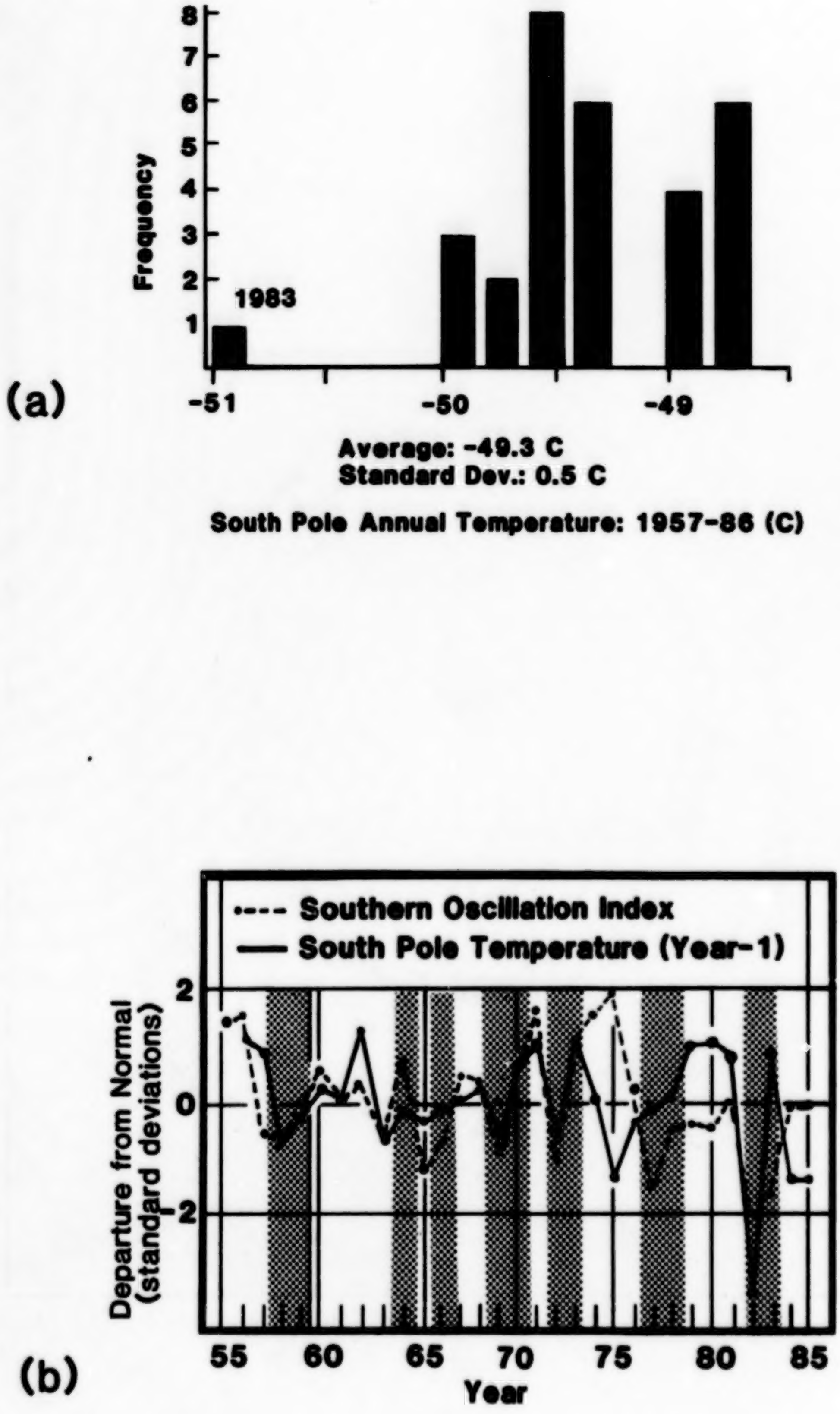

(a)
Frequency
1
2
3
4
5
6
7
8
1983
-51
-50
-49
Average: -49.3 C
Standard Dev.: 0.5 C
South Pole Annual Temperature: 1957-86 (C)
(b)
Southern Oscillation Index
South Pole Temperature (Year-1)
Departure from Normal (standard deviations)
2
0
-2
55
60
65
70
75
80
85
Year

Figure 3.

BLANK

PAGE

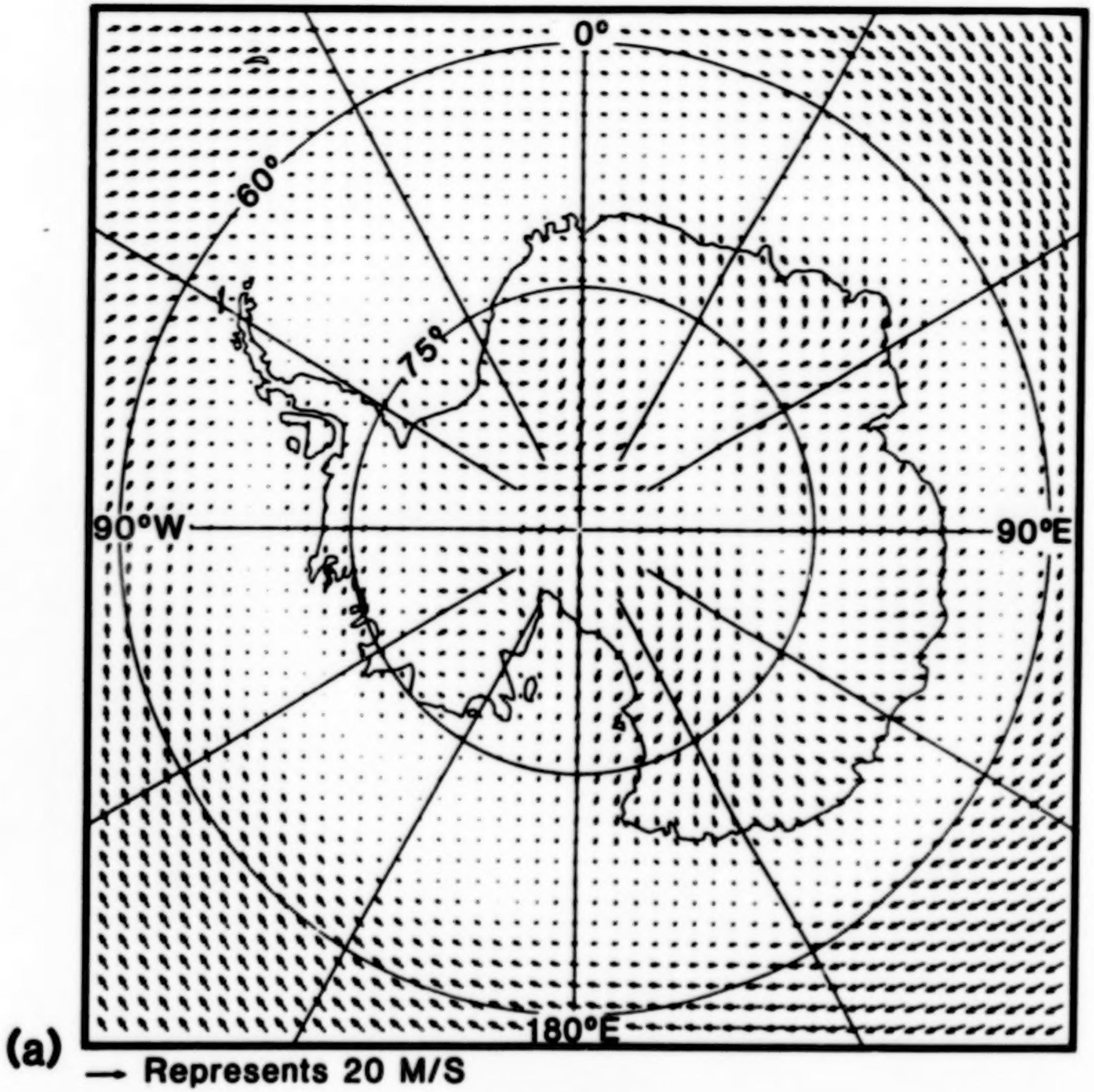


Figure 4.

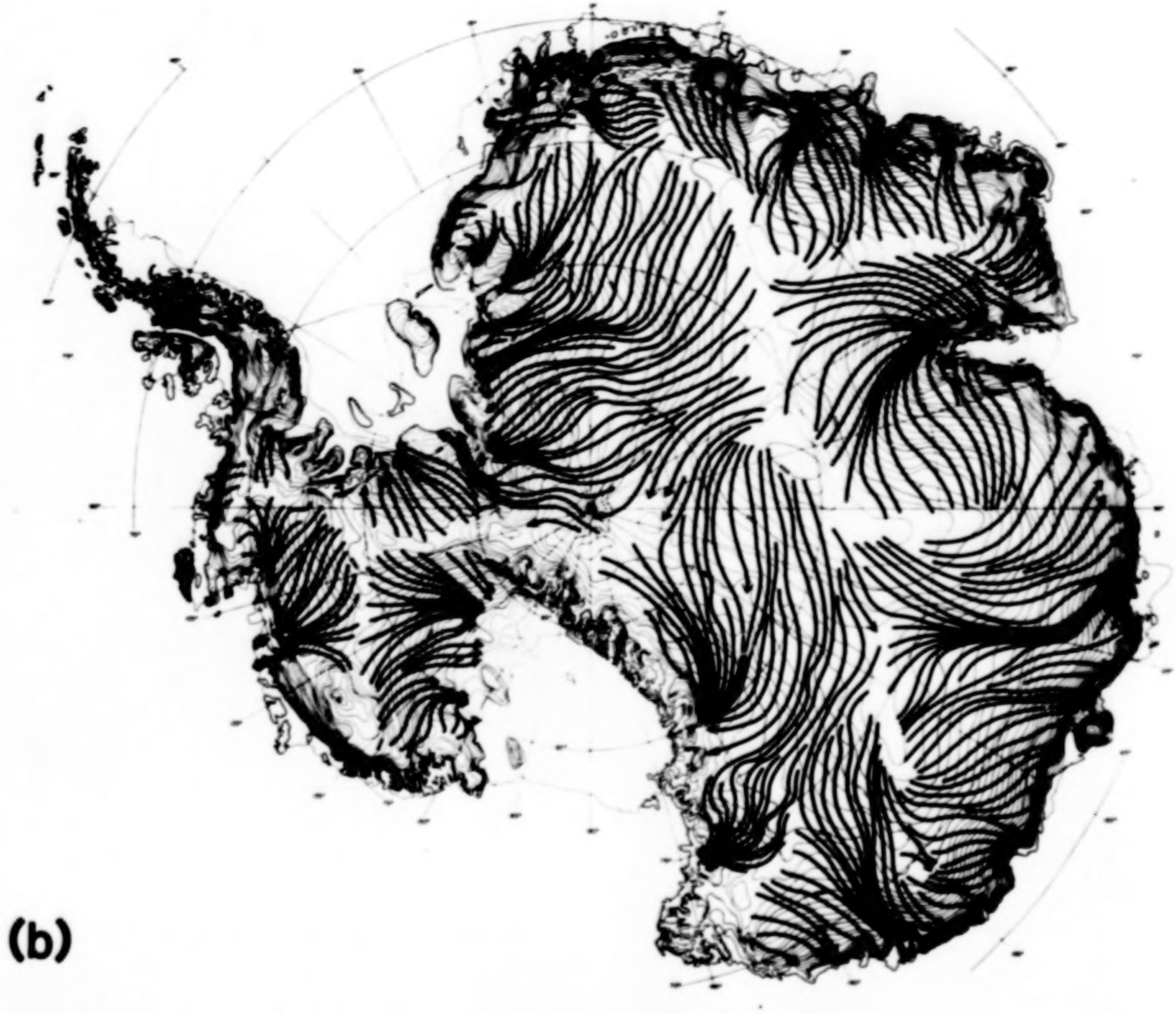


Figure 4 (continued).

Sea-Level Response to Ice Sheet Evolution: An Ocean Perspective

Stanley S. Jacobs

Lamont-Doherty Geological Observatory
of Columbia University
Palisades, NY 10964

Abstract

The ocean's influence upon and response to Antarctic ice sheet changes is considered in relation to sea level rise over recent and future decades. Assuming present-day ice fronts are in approximate equilibrium, a preliminary budget for the ice sheet is estimated from accumulation vs. iceberg calving and the basal melting that occurs beneath floating ice shelves. Iceberg calving is derived from the volume of large bergs identified and tracked by the Navy/NOAA Joint Ice Center and from shipboard observations. Basal melting exceeds 600 km^3yr^{-1} and is concentrated near the ice fronts and ice shelf grounding lines. An apparent negative mass balance for the Antarctic ice sheet may result from an anomalous calving rate during the past decade, but there are large uncertainties associated with all components of the ice budget. The results from general circulation models are noted in the context of projected precipitation increases and ocean temperature changes on and near the continent. An ocean research program that could help refine budget estimates is consistent with goals of the West Antarctic Ice Sheet Initiative.

Introduction

The ocean both forces and responds to ice sheet evolution. It provides the evaporative moisture that nourishes the ice sheet and regulates the sources of that moisture by its waxing and waning fields of sea ice. The ocean alters the position of ice sheet grounding lines and the dimensions of ice shelves by changes in sea level, melting, freezing and mechanical attrition. It exports calved icebergs in the strong coastal currents and polar gyres, and accepts surface runoff and subsurface glacial meltwater in its top and bottom layers. It damps the potential rate of climate change by its large heat capacity, deep vertical mixing in the polar regions, heat transfers associated with melting and freezing, and by its uptake of greenhouse gases. The ocean may also have discrete stable modes of circulation that promote, retard or respond in different ways to ice sheet growth. More than 90% of the earth's ice sheet lies on the Antarctic continent.

From paleoclimate records, it is known that global sea level has risen at an average rate of ~6mm yr^{-1} over the past 20,000 years, including rapid increases that exceeded 24 mm yr^{-1} (Fairbanks 1989). A slower rise over the past century has been attributed in part to an ocean volume increase resulting from a parallel rise in air temperature, and to the retreat of temperate glaciers (Gornitz et al. 1982; Meier 1984). The remainder is unaccounted for or ascribed to changes in surface and groundwater reservoirs and to uncertainties in mass balance of the large ice sheets (Meier 1990a). This balance is unknown, but the Antarctic portion is generally considered to be slightly positive or close to a state of equilibrium (Budd and Smith 1985; Bentley 1989), while Greenland's contribution to sea level could be of either sign (Reeh 1985; Zwally 1989; Braithwaite and Olesen 1990). The slow response time of glacial ice complicates the problem of determinining ice sheet mass balance over the short term, as does the lack of good-quality data, the low signal to noise ratio, and the inherent difficulty of making field measurements in the polar regions.

The average rate of global sea level rise over the past century has been 1-2 mm yr^{-1}, with higher values (2.3 - 2.4 mm yr^{-1}) over the past 50 years supported by glacial isostatic adjustments (Peltier and Tushingham 1989; Barnett 1990). Taking 1.5 mm yr^{-1} as a working number, this addition to the ~360 x 10^6 km^2 ocean surface would equal 540 km^3 yr^{-1} of water, or ~635 km^3 yr^{-1} of ice. For comparison, that volume is roughly one quarter of the present-day annual accumulation on the Antarctic ice sheet, or the approximate annual flow of the Mississippi River. Taking the ice-ocean boundary as a reference interface, this background paper will focus upon the attrition side of Antarctica's ice budget, primarily calving and basal mass balance. Freezing and melting beneath ice shelves will be emphasized because these processes are often discounted, are involved in deep ocean ventilation and may be sensitive to short-term climate change. Sub-ice shelf and general circulation models will be discussed in the context of basal fluxes and possible future ocean changes near Antarctica. Finally, an outline will be given for the oceanographic components of a potential multidisciplinary program to study the relationship of the Antarctic ice sheet to global climate change.

Iceberg Calving

The calving of icebergs from Antarctica is the largest factor in the attrition of the ice sheet, defined here as extending to the ice front perimeter. Calving from floating ice shelves has no immediate impact upon sea level, but reduced back-pressure on the grounded ice should permit faster seaward flow of ice streams across the grounding lines (Hughes 1973). From relatively small changes in the average positions of ice fronts over the past several decades (Markov et al. 1968; Barkov 1985; Zakarov 1988), it can be argued that a rough balance exists between ice sheet accumulation and wastage. However, outflow may not respond rapidly to changes in accumulation, and is likely to be balanced over time by adjustments in calving rate. Similarly, an ice shelf might thicken or thin for some time before major perturbations appeared in its equilibrium calving line. A large part of the attrition may occur by major calving events at periods of several decades, but early ice front positions were often based upon navigational records of low precision, and little of the Antarctic perimeter or its ice velocity has been systematically monitored. Satellite techniques for mapping ice fronts have shown considerable promise (Thomas et al. 1983; Zwally et al. 1987)s and new sensors expected to be in polar orbit late in this decade should allow routine monitoring to begin. Achieving significant results over a short period of time will be difficult, but good spatial and temporal coverage of the ice fronts, ice thickness and ice velocity should aid in the development of ice dynamics and calving models.

Attempts to assess the attrition of Antarctica by iceberg volumetrics from ships of opportunity have not generally been held in the highest regard. The uncertainties of estimating iceberg sizes, ocean area surveyed and typical iceberg lifetimes are compounded by the difficulties of accounting for duplicate sightings, seasonal and areal variability, and wastage between the time of calving and observation. The report "Glaciers, Ice Sheets and Sea-Level" (NRC/DOE 1985) discounted its only relevant contribution on this topic, wherein Orheim (1985) suggested a negative Antarctic ice sheet mass balance based upon a 1977-1984 calving estimate of 2.3×10^{15} kg (~2700 $km^3\ yr^{-1}$). That study covered a period when the major ice shelf fronts were known to be advancing (Lange & Kohnen 1985; Jacobs et al. 1986) and thus could not have contributed much to the calculated volume. Although the Amery, Filchner-Ronne and Ross Ice Shelves drain more than half of the grounded ice sheet, much of that is the low-precipitation continental interior, so these drainage systems account for only one third of the outflow (Giovinetto and Bentley 1985). Given the high sensitivity of the derived production rate to rough estimates of iceberg lifetimes, a remarkable feature of the Orheim (1985) result and earlier estimates in the same range may be their close approximation to ice sheet accumulation values. Nonetheless, it is worth re-examining the iceberg data in the light of more recent observations.

Much useful information about iceberg distributions, sizes and freeboards can be derived from comprehensive surveys. By extrapolation from a smaller count, the annual number of bergs in the Southern Ocean can apparently exceed 300,000, including all bergy bits >10 m (Orheim 1985). That implies one chunk of ice for each 120 km^2 of the 36 x 10^6 km^2 ocean surface south of the Polar Front, consistent with fig. 8 in Weeks and Mellor (1978). Wadhams (1988) calculated one iceberg for each 164 km^2 of ocean surface for a short observational period in a restricted area of the winter South Atlantic, where survival time was estimated to be < 6 months and it was difficult to detect bergs smaller than 115 m in diameter. Annual production rates inferred from iceberg distributions are inversely proportional to their lifetimes, which Orheim (1985) took to be 4 years. Survival times are longer for icebergs near Antarctica, berg dimensions are larger within the pack ice and numbers of bergs decrease away from the coastline (Orheim 1985; Hult and Ostrander 1973; Morgan and Budd 1978). Grounding is common on bottom shoals of the continental shelf, while larger waves, higher temperatures, stronger currents and less sea ice promote more rapid mechanical and thermal disintegration over the deep ocean. These data suggest that volumetric estimates might be refined by distinguishing between iceberg populations in different areas or size categories.

Orheim (1989) indicated that for the previous several years the annual calving rate of small icebergs (main axis <22 km) was approximately constant both in numbers and total mass, and exceeded the annual mean mass of large icebergs (>22 km) calved during the same period. This covered a period (1986-87) of major iceberg calving in the Weddell and Ross Seas, but we know that large calving events also spawn numerous smaller bergs and that different size classes experience different fates (Keys et al. 1990). For more than a decade, large icebergs (> 28 km) have been identified and tracked from satellite images by the Navy-NOAA Joint Ice Center (e.g., NPOC 1987-88). The annual volume of newly-calved icebergs has been tabulated from their 1979-1989 data in Fig. 1, with the area of each iceberg conservatively approximated by an ellipse. Exceptions are the 1986 Filchner and 1987 Ross icebergs, where we have adopted the Landsat-derived areas of Ferrigno and Gould (1987) and Keys et al. (1990). Average thickness was taken to be 250 m, close to that estimated for the Ross (B-9) iceberg, which was probably thicker than the Larsen calf and thinner than the Filchner ones. From 1982-88, the average large berg volume was ~1520 km^3 yr^{-1}, but for the full 11-year period in Fig. 1 the average was 1279 km^3 yr^{-1}. Depending upon their size limits (<1 km or <28 km) we could get from 839 to 1279 km^3 yr^{-1} for small bergs. Approximating a small berg value from the average of this range (1059 km^3 yr^{-1}) and adding the 1979-89 large berg average, then total calving equals 2338 km^3 yr^{-1}. This is again close to the accumulation value, but we still have to consider basal melting.

Basal Mass Balance

Basal melting of the ice shelves is sometimes ignored in mass balance estimates (e.g., Budd and Smith 1985), for a variety of reasons. It is hidden beneath the ice or sea surface, occurs at temperatures below 0°C, is believed to be negligible, contrary to or balanced by observed basal freezing, or it is seaward of the grounding line and can thus have no impact on sea level. It is true that melting or freezing at the base of floating ice does not appreciably alter sea level, but most compilations of precipitation on Antarctica include the ice shelves. The total accumulation of Giovinetto and Bentley (1985) drops from 1962.7×10^{12} kg yr^{-1} (~2310 km^3 yr^{-1} at an ice density of 0.85) to ~1730 km^3 yr^{-1} when ice shelves, ice rises and ice islands are excluded from the accounting. With sufficient satellite data it may be possible at some time in the future to define all grounding lines and monitor the ice flow across them. In the interim it is more practical to include the ice shelves, and more important to investigate their influence upon mass flux off the continent.

Ocean data and models generally point to net basal melting beneath the ice shelves (e.g., Jacobs et al. 1979; Robin et al. 1983; Potter et al. 1988; Hellmer and Olbers 1989). These results might appear contradicted by measured or inferred sea ice accumulations on the ice shelf bases (e.g., Morgan 1972; Neal 1979; Zotikov et al. 1980; Engelhardt and Determann 1987). However, most ice cores and holes have been centrally located on the large ice shelves where most models also show net freezing. Thick sea ice deposits on the large ice shelf bases may be accounted for by an "ice-pump" (Lewis and Perkin 1986), whereby ice is removed from some regions and redeposited in others. This process has important implications for water column properties, basal mass balance and the stability of ice shelf pinning points. In the discussion that follows, basal mass balance will include both melting and freezing.

For several decades, investigators have reasoned from measurements near the major ice shelves that melting in the sub-ice cavities could be on the order of tens of cm yr^{-1} (Wexler 1960; Crary 1961; Thomas and Coslett 1970; Robin 1979). This inference was predicated upon an active circulation beneath the ice shelves, coupled with an oceanic heat source on or north of the continental shelf or a temperature differential afforded by a decrease of the *in situ* freezing point with pressure (Doake 1976; Jacobs et al. 1979). Numerical models subsequently showed that the energy for vertical mixing under the ice, essential to bring heat into the near-freezing boundary layer, could be derived from the tidal and thermohaline circulations (MacAyeal 1984a; 1985). Gravity and direct current measurements documented the active tidal circulations, and closely-spaced hydrographic observations revealed the limited dimensions of inflowing and outflowing shelf waters.

Long-term instrument arrays near the Ross Ice Shelf front in the mid-1980's focused upon a subsurface region of relatively 'warm' water, ~1°C above the in situ freezing point in the austral summer, apparently derived from the continental slope region more than 250 km to the north (fig. 6b in Jacobs et al. 1985). It was hypothesized that this water could supply sufficient heat to melt ~40 km^3 yr^{-1} off the ice shelf base for each cm s^{-1} of net southward flow. Current and temperature records confirmed a persistent southward flow of this water during 1983, averaging ~5 cm s^{-1}. In spite of a seasonal temperature signal, some of the warmest temperatures appeared during the austral winter (Pillsbury and Jacobs 1985). However, a more closely-spaced array during 1984 showed that a major fraction of this ocean heat was recirculated to the open sea a few tens of km to the west of the inflow (Fig. 2). It is not yet known whether similar circulation patterns exist where warm water is drawn toward other ice shelves and glacier tongues, as illustrated by Foldvik et al. (1985) and Jacobs (1989).

The concept of an active circulation and net melting beneath the ice shelves has also been supported by geochemical measurements, from which it is possible to identify the distinctive properties of glacial meltwater, or to calculate the residence time of seawater beneath the ice (Michel et al. 1979; Jacobs et al. 1985; Potter and Paren 1985; Schlosser 1986; Fahrbach et al. 1991; Trumbore et al. 1991). Utilizing a shelf water chlorofluorocarbon (CFC) model, the latter study indicated that the time for High Salinity Shelf Water to evolve into Ice Shelf Water (ISW, with $T<T_{frs}$, i.e., temperatures below the surface freezing point) beneath the Ross Ice Shelf could be as little as 3.5 years. Ocean water properties have also been used to establish the presence of glacial meltwater in Antarctic Bottom Water (Weiss et al. 1979; Foldvik and Gammelsrod 1988; Schlosser et al. 1990). This finding has been extended by some investigators, who believe that ISW is primarily responsible for the coldest bottom water currently observed in the Weddell Sea (Foldvik et al. 1985; Jacobs 1986).

Recent models of the sub-ice shelf circulation and ice shelf mass balance have displayed a number of common features. These include high melting near the grounding lines, low melting or sea ice accumulation over wide areas of the ice shelf base, and high melting near the ice fronts. Integrating a Filchner Ice Shelf transect in the thermohaline circulation model of Hellmer and Olbers (1989) from the grounding line to the northern edge of the accumulation zone (Fig. 3) results in an average melting of ~45 cm yr^{-1}. Similar treatment of a Ronne Ice Shelf transect from the Rutford Ice Stream (fig. 10 in Jenkins and Doake 1991) gives ~65 cm yr^{-1} melting for the region more than 100 km south of the ice front, including a 200 km section where net accumulation prevails. Due to the pressure dependency of *in situ* freezing temperature, the melting rate is directly proportional to depth, leading to > 4 m yr^{-1} melting in the vicinity of the relatively deep Rutford transect grounding line (Jenkins and Doake 1991; Pozdeyev & Kurinen 1987). However, the Ronne Ice Shelf includes a more extensive area of basal freezing east of the Rutford transect (Engelhardt and Determann 1987). For these reasons, 45 cm yr^{-1} will be used as a working value for the entire Filchner-Ronne Ice Shelf region > 100 km south of the ice front, yielding 181 km^3 yr^{-1}. This rate is lower that an estimate of 211-246 km^3 yr^{-1} that can be derived from hydrochemical data in Schlosser et al. (1990), assuming 6-7$^{o}/_{oo}$ meltwater in an ISW outflow of 10^6 m^3 s^{-1} (Foldvik et al. 1985). That higher range may indicate other sources of meltwater, some perhaps in the inflow (Weiss et al. 1979), or that the model-derived estimate is overly conservative.

Glaciological models for the region well south of the Ross Ice Shelf front have shown equilibrium net melting on the order of 12-17 cm yr^{-1} (Shabtaie and Bentley 1987; Lingle et al. 1990). Oceanographic data and models have suggested similar to higher rates (Jacobs et al. 1979; Scheduikat and Olbers 1990), some of which appear to be at odds with a sea ice core retrieved from the central ice shelf (Zotikov et al. 1980). However, the active ocean circulation beneath an ice shelf means that the products of freezing and melting are dispersed far from their origins. The primary large-scale outflows appear as Ice Shelf Water temperature minima, observed, e.g., in vertical ocean property profiles near the front of the Filchner Ice Shelf. These features are reproduced by the Hellmer and Olbers (1989) model, from which their depth is shown to be dependent upon the thermohaline characteristics of inflowing High Salinity Shelf Water. Perhaps for this reason and the accompanying changes in circulation strength and melt rate, temperature minima within the Ross Sea ISW outflow tend to cluster at discrete depth intervals, as illustrated by the shaded bands in Fig. 4. Interannual variability in the production and volume of High Salinity Shelf Water or multiple circulation cells beneath the ice shelf could also influence the outflow characteristics.

The salinity difference between High Salinity Shelf Water inflow and the ISW outflow can be combined with current measurements to estimate a basal melt rate for the Ross Ice Shelf. Apparent melting exceeds the heat available, judging by the observed temperature change between High Salinity Shelf Water and ISW. This discrepancy may be accounted for by incorporation of outflow from beneath the grounded ice sheet, by ice crystal formation in the near-freezing water column or by tidal mixing near the ice front (Foldvik and Kvinge 1974; MacAyeal 1984a; Dieckmann et al. 1986). An estimate for salinity at the time of inflow can be obtained from a 1982 profile in fig. 4.1 of Jacobs (1985). Integrating over the region where $T<T_{frs}$ in Fig. 4 and using the long-term average velocity (1.9 cm s^{-1}) of 4 current meters sited as shown, the meltwater in this plume is then ~79 km^3 yr^{-1}. Removing an equivalent volume of ice from the area of the ice shelf base >100 km south of the ice front would correspond to a melt rate of ~20 cm yr^{-1}, or ~88 km^3 yr^{-1}. That is roughly one half the Filchner-Ronne melt rate and may be consistent with less bottom water production in the Ross Sea, which St. Pierre (1989) has attributed to differences between the tidal regimes. It does not include meltwater in the circulation cell responsible for the lower-salinity boundary layer beneath the ice shelf at J-9 (Jacobs et al. 1979).

The George VI Ice Shelf has been accorded a 2.1 m yr^{-1} melt rate (53 km^3 yr^{-1}) from a study of glaciological and oceanographic data (Potter and Paren 1985). Net basal freezing of -0.6 m yr^{-1} along the central flow line of the Amery Ice Shelf has been calculated by Budd et al. (1982). Since oceanographic measurements near the Amery ice front (Smith et al. 1984) also suggest basal melting, that freezing rate has tentatively been applied to only half of the ice shelf area, resulting in -12 km^3 yr^{-1} (Table 1).

Models and indirect observations suggest high basal melt rates near the ice fronts (references on p. 77 of Jacobs et al. 1985; Fahrbach et al. 1991). These results are consistent with the low temperature and salinity in oceanographic transects perpendicular to the coastlines (e.g., fig. 5 in Jacobs 1989). For the eastern Ross Ice Shelf front, Thomas and MacAyeal (1982) estimated a basal melt rate of 0.7 m yr^{-1}, decreasing to near zero 100 km south of the front. That conservative estimate corresponds to ~3.5 km^3 yr^{-1} for each 100 km of ocean frontage. Excluding the interior Filchner-Ronne, Ross, George VI and Amery Ice Shelves which are covered separately above, most of the remaining 648 x 10^3 km^2 ice shelf area (from Drewry 1983), falls within the 100 km coastal band and nets 227 km^3 yr^{-1}. Laboratory experiments indicate that wall melting exceeds basal melting by a factor of ten (Neshyba and Josberger 1980), which would be ~350 cm yr^{-1} in this case. With a coastline of 31,876 km of which 57% is ice shelves, outlet glaciers and ice streams (Drewry 1983), we obtain another 16 km^3 if the average wall thickness is 250 m.

To the estimates above must be added the small contributions from surface runoff and outflow from beneath the grounded ice sheet. Robin (1987) estimated surface runoff of 36 km^3 yr^{-1} over that 2% of the ice sheet in the ablation area . Melting beneath the grounded ice sheet from geothermal heat flux and glacier sliding may be two orders of magnitude less than beneath the ice shelves and be applicable to half the grounded area (Zotikov, 1963; Budd & Jenssen 1987; Engelhardt et al. 1990). In this case 21 km^3 yr^{-1} will be released along the grounding lines. The sum of all in situ melting components is 610 km^3 yr^{-1} (Table 1), higher than most earlier estimates (e.g., p. 216 in Barkov 1985), but consistent with the Jacobs et al. (1985) appraisal from limited oxygen isotope data.

The model results and data-based mass balances are not well constrained due to uncertainties in the underlying assumptions, scarcity of data, and high spatial and seasonal variability in the coastal ocean. Nonetheless, these conservative estimates of the net basal melting (~610 km^3 yr^{-1}, from Table 1) added to the iceberg volume (2338 $km^3\ yr^{-1}$, from above) indicate that the Antarctic ice sheet was in a negative state of balance for the 1979-89 period. The difference between attrition (2948 $km^3\ yr^{-1}$) and accumulation (2535 $km^3\ yr^{-1}$) is equivalent to a sea level rise of ~1 mm yr^{-1}. A larger difference could result from several other plausible interpretations of the data. While the uncertainties are no less than the net result, the most probable value lies on the negative side of balance. Aside from illustrating the large basal meltwater component, this analysis suggests that we cannot rely upon Antarctica to be a restraining influence on future sea-level rise (NRC/DOE 1985; Bentley 1989; Meier 1990b).

Present Variability and Future Change

The large-iceberg calving rate (Fig. 1) is characterized by a high spatial and temporal variability. We can expect less interannual variability in the basal melt rate, given a steady-state ocean circulation that is buffered near the coastline by large volumes of seawater formed at the sea surface freezing temperature (MacAyeal 1984b). However, some data show large interannual salinity changes in high salinity shelf water at the same site and time of year (0.1-0.15$^o/_{oo}$ in fig. 4.1 of Jacobs 1985). These observations are not from an atypical location, as transects along the full Ross Ice Shelf front at ~8 yr intervals beginning in 1967-68 show a shift to lower salinities and higher temperatures. This does not necessarily indicate progressively more basal melting or less sea ice formation, but does reveal considerable natural variability in the shelf water mass that is important to ISW and bottom water production. Significant annual changes in the properties of bottom water have been observed in transects of salinity and silicate across the continental margin in the northwest Weddell Sea (fig. 2 in Foster and Middleton 1979), possibly related to fluctuations in the characteristics or formation rate of shelf waters. As yet there is little evidence of major variability in the winter atmospheric forcing on the Antarctic continental shelf, but the annual duration of ice cover is known to differ by several weeks in the Ross Sea (Jacobs and Comiso 1989). At present we have only a qualitative understanding of how sea ice cover, thickness and transport relates to brine drainage on the shelf and to the conversion of shelf waters into ISW and bottom water.

The strength of the ocean circulation beneath ice shelves may be highly sensitive to small changes in the properties of inflowing seawater and to the shape of the sub-ice cavity. Hellmer and Olbers (1989) varied input temperature and salinity by .02°C and .02$^o/_{oo}$ and found significant differences between the rates and zones of basal melting and freezing on the Filchner Ice Shelf base (Fig. 3). That variability is small relative to the interannual shifts and seasonal cycles that have been measured along the Ross Ice Shelf front. One sensitive aspect of the cavity shape is the depth of the grounding line, where melting may be proportional to the freezing point depression. This has interesting positive feedback implications if West Antarctic ice sheet stability is linked to grounding line retreat into the deeper interior basins (Hughes 1973). Glaciological models with ocean/ice sheet interactions also suggest that basal mass balance will be sensitive to ocean temperature changes. Most probable bounds, assuming ice shelf thinning of 10% or a basal melting increase of ~1 m yr^{-1}, would contribute 4-30 cm to sea level rise by the year 2100 (Lingle 1985; Thomas 1987). Lingle et al. (1990) show rapid and sustained thinning of the Ross Ice Shelf when the rate of basal melting is increased linearly over 150 years to 2 m yr^{-1}. That is no more than the present-day melt rate of the George VI Ice Shelf (Potter and Paren 1985) where nearly undiluted 'warm' CDW floods the continental shelf. At most other locations CDW access to the ice shelf bases appears to be limited by the presence of higher density shelf water produced by sea ice freezing or lower density shelf water formed by ice melting into westward-flowing coastal waters (Jacobs et al. 1985; Fahrbach et al. 1991). It may also be limited to some extent by the well-developed oceanic frontal zone over the upper continental slope (Jacobs 1989).

Could climate changes cause a shift between a low-melting mode (tens of cm yr^{-1}) associated with High Salinity Shelf Water drainage into the sub-ice cavities, and a high-melting mode (m yr^{-1}) associated with deep water (CDW) intrusions onto the shelf? Warming scenarios linking decreased sea ice formation with less dense shelf water would permit more CDW onto the continental shelf (Parkinson and Bindschadler 1984; Lingle 1985). If precipitation also increased, CDW temperatures could warm by ~0.5°C (Gordon 1983). This would result from a greater density contrast between a warmer or fresher surface layer and the deep waters, perhaps abetted by lower wind speeds lessening the divergence and replacement of the ice and mixed layer. The shift would be toward a more stable Arctic-type pycnocline, through which the vertical heat flux is an order of magnitude less than in the Southern Ocean. Gordon (1982) has used Weddell Sea data to demonstrate the short-term destruction of a marginally stable pycnocline, with widespread impacts on the sea ice cover and deep water properties. CDW temperature increases could be accompanied by greater heat transport across the oceanic slope front and onto the continental shelf. At present there is a direct proportionality between the temperatures of deep water and 'warm' shelf intrusions in the Ross and Weddell Seas, but not at some other shelf locations.

A precipitation change of 3-4 cm yr^{-1}, ~25% above current levels, has been predicted from general circulation models (GCM's) and from accumulation changes on Antarctica since the last glacial maximum (Manabe et al. 1990; Robin 1987). The former derives from model results that show enhanced greenhouse-driven atmospheric warming in the polar regions, particularly during the winter season. Ocean areas near Antarctica have been cited as one of the regions where unambiguous warming would appear earliest (Hansen et al. 1988). The estimated amplitude of surface warming has varied with location around the continent and the interannual variability may be high. In one model a subsurface maximum in ocean warming appeared only near the continental shelf break in the potentially sensitive Amundsen Sea region (Schlesinger 1985). More recent NOAA/GFDL model results have shown an asymmetric response to climate warming, with less of an air temperature increase in the Southern Hemisphere (Bryan et al. 1988; Stouffer et al. 1989; Manabe et al. 1990). This is attributed to the large thermal inertia and greater convective overturning in this ocean-dominated hemisphere, and to the presence of the Antarctic Circumpolar Current. Increased precipitation in high latitudes results in a stronger pycnocline, reduced mixing between the surface layer and CDW, and a decrease in sea surface temperature during the latter part of a 60-year simulation. Surface cooling centered in the Amundsen Sea and higher deep water temperatures are also consistent with the discussion above, but a narrow zonal band of >0.5°C warmer water to depths >2000 m near the Antarctic continent (fig. 2 in Stouffer et al. 1989) may be anomalous. Unless the ocean circulation were to change significantly, that water would be regionally removed northward as bottom water or upwelled into the surface layers and onto the continental shelf. In the latter case, the warmer subsurface shelf waters might increase basal melting (MacAyeal 1984b), a possible counterbalance to more precipitation on the continent.

What next?

The apparent negative mass balance derived above for the Antarctic ice sheet is consistent with the observed direction of sea level change, and does not exceed recent estimates of 2.3 - 2.4 mm yr^{-1} when other contributions are added. It is contrary to recent estimates of Antarctica's negative contribution to sea level rise. The result must be considered preliminary and may be due to short-term variability of iceberg calving and the ocean circulation, the general scarcity of field data, overly primitive models, or the underestimation of present-day accumulation on the ice sheet. The uncertainties associated with these factors could be reduced and some valuable baselines established by a well-focused program of ocean and ice sheet measurements and modeling (Bindschadler et al. 1990). The oceanographic goals of that program should include:

1. Direct measurements of Ice Shelf Water generation and outflow, its spatial and temporal variability, and its fate in the deep ocean. Observations must profile the full water column over several annual cycles and include a mix of geochemical tracers that can yield residence times over relevant time and space scales.

2. Direct sampling and measurement of seawater beneath a major ice shelf by means of holes sited along transects from the ice shelf front to and beyond typical grounding lines. Multiple ice holes and ice cores are essential to resolve regional variability of melting and freezing, particularly near grounding lines, basal crevasses, ice rises and ice fronts. Simultaneous time-

series measurements are needed of the ice shelf thickness, velocity, accumulation, and basal mass balance.

3. Icebreaker penetration and detailed oceanographic sampling of the anomalous and largely unknown Amundsen-Bellingshausen continental shelf region.

4. Modeling of the continental shelf and sub-ice shelf ocean circulations. Models must be sensitized to the measured ranges of temperature, salinity, currents and chemical tracers, and should focus upon regions near the grounding lines, ice fronts and shelf break that may be more highly sensitive to climate change.

5. Establishment of a baseline ice sheet perimeter and elevation, by GPS-controlled shipboard navigation and radar if necessary, against which future satellite mapping may be compared. We assume that polar orbiting satellites will carry altimeters capable of measuring ice extent and thickness within the next decade, but there will be a need for ground-truthing and the siting of weather stations and transmitters for ice velocity measurements.

A schematic figure for portions of such a program (Fig. 6) focuses upon the Ross Sea and Ice Shelf because of its more extensive data base and logistic accessibility. With the AWS/AVHRR/SAR weather/satellite facilities existing or planned for McMurdo, the Ross Sea polynyas can be closely monitored for response to atmospheric forcing. Sediment cores could be obtained at each site, with ice cores and precipitation variability measurements along and across flowlines. Automatic Weather Station or Argos transmitters would provide atmospheric and velocity data between site visits. The proposed ISW transect would cross the Ross Ice Shelf front near its northwest corner, now farther north than at any time in the past 150 years (Jacobs et al. 1986). It could thus provide the opportunity to study the dynamics of an old ice front, and/or a new one, during the term of the project.

Acknowledgments

Discussions with C. Doake, E. Domack, H. Hellmer, A, Jenkins, D. MacAyeal, M. Meier and colleagues at the NASA SeaRISE Workshops were helpful in preparing this paper. Support was derived from National Science Foundation grants DPP-85-02386 and 90-08906. Contribution # 4709 of the Lamont-Doherty Geological Observatory.

References

Barkov, N.I. (1985) Ice Shelves of Antarctica. E.S. Korotkevich [Ed.], Amerind Publ. Co., New Delhi & New York, 262 pp. [Translation of Shel'fovye ledniki Antarktidy, 1971].

Barnett, T.P. (1990) Recent Changes in Sea Level: A Summary. In: Sea-Level Change, Studies in Geophysics, Nat'l Res. Council, Nat'l Acad. Press, Wash., DC, 37-51.

Bentley, C.R. (1989) The current mass balance of the Antarctic Ice Sheet. Eos, Transactions, Am. Geophys. Un., 70(43), 1002 (abstract).

Bindschadler, R. et al. (1990) SeaRISE: A Multidisciplinary Reaearch Initiative to Predict Rapid Changes in Global Sea Level Caused by Collapse of Marine Ice Sheets. Proc of a Workshop, College Park, MD, 23-25 Jan. 1990, NASA Conference Publication 3075, 47pp.

Braithwaite, R.J. & O.B. Olesen (1990) Increased ablation at the margin of the Greenland ice sheet under a greenhouse-effect climate. Ann. of Glac., 14, 20-22.

Bryan, K., S. Manabe & M.J. Spelman (1988) Interhemispheric asymmetry in the transient response of a coupled ocean-atmosphere model to a CO_2 forcing. J. Phys. Oceanogr., 18, 851-867.

Budd, W.F., M.J. Corry & T.H. Jacka (1982) Results from the Amery Ice Shelf Project. Ann. Glaciol., 3, 36-41.

Budd, W.F. & I.N. Smith (1985) The state of balance of the Antarctic Ice Sheet - An Updated Assessment 1984. In: NRC/DOE (1985) Glaciers, ice sheets, and sea level: Effect of a CO_2-induced climatic change, Report of a workshop, Seattle, 13-15 Sept. 1984, Nat'l. Res. Council, Polar Res. Bd., Dept. of Energy (DOE/EV/60235-1), 172-177.

Budd, W.F., & D. Jenssen (1987) Numerical modeling of the large-scale basal water flux under the West Antarctic ice sheet. In: C.J. van der Veen & J. Oerlemans (Eds), Dynamics of the West Antarctic Ice Sheet, D. Reidel, Boston, 293-320.

Crary, A.P. (1961) Glaciological regime at Little America Station, Antarctica. Jour. Geophys. Res., 66, 871-878.

Dieckmann, G., G. Rohardt, H. Hellmer & J. Kipfstuhl (1986) The occurrence of ice platelets at 250m depth near the Filchner Ice Shelf and its significance for sea ice biology. Deep-Sea Res., 33(2), 141-148.

Doake, C.S.M. (1976) Thermodynamics of the interaction between ice shelves and the sea. Polar Record, 18(112), 37-41.

Doake. C.S.M. (1985) Glaciological evidence: Antarctic Peninsula, Weddell Sea. In: NRC/DOE (1985) Glaciers, ice sheets, and sea level: Effect of a CO_2-induced climatic change, Report of a workshop, Seattle, 13-15 Sept. 1984, Nat'l. Res. Council, Polar Res. Bd., Dept. of Energy (DOE/EV/60235-1), 197-209.

Drewry, D. (Compiler, 1983) Antarctica: Glaciological and Geophysical Folio. Scott Polar Res. Inst., Cambridge.

Engelhardt, H. & J. Determann (1987) Borehole evidence for a thick layer of basal ice in the central Ronne Ice Shelf. Nature, 327(6120), 318-319.

Engelhardt, H., N. Humphrey, B. Kamb & M. Fahnestock (1990) Physical conditions at the base of a fast moving Antarctic ice stream. Science, 248, 57-59.

Fahrbach, E., G. Rohardt, P. Schlosser & R. Bayer (1991) Supression of bottom water formation on the Southeast Weddell Sea shelf due to melting of glacial ice. ms in preparation.

Fairbanks, R.G. (1989) A 17,000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. Nature, 342, 637-642.

Ferrigno, J.G. & W. Gould (1987) Substantial changes in the coastline of Antarctica revealed by satellite imagery. Polar Record, 23, 577-583.

Foldvik, A. & T. Kvinge (1974) Conditional instability of sea water at the freezing point. Deep-Sea Res., 21(3), 169-174.

Foldvik, A., T. Gammelsrod & T. Torresen (1985) Circulation and water masses on the southern Weddell Sea shelf. In: Oceanology of the Antarctic Continental Shelf, Ant. Res. Ser., 43, Am. Geophys. Un., Wash. D.C., pp. 5-20.

Foldvik, A., & T. Gammelsrod (1988) Notes on Southern Ocean hydrography, sea ice and bottom water formation. Paleo. Paleo. Paleo., 67, 3-17.

Foster, T.D. & J.H. Middleton (1979) Variability in the bottom water of the Weddell Sea. Deep-Sea Res., 26A, 743-762.

Giovinetto, M.B. & C.R. Bentley (1985) Surface balance in ice drainage systems of Antarctica. Ant. Jour. of the U.S., 20(4), 6-13.

Gordon, A.L. (1982) Weddell deep water variability. J. Mar. Res. 40(Suppl.), 199-217.

Gordon, A.L. (1983) Comments about the ocean role in the Antarctic glacial ice balance. Proceedings: CO_2, Science and Concensus, 19-23 Sept., 1982, Berkeley Springs, W. Va.

Gornitz, V., L. Lebedeff and J. Hansen (1982) Global sea level trend in the past century. Science, 215, 1611-1614.

Hansen, J., I. Fung, A. Lacis, D. Rind, S. Lebedeff, R. Ruedy & G. Russell (1988) Global climate changes as forecast by Goddard Institute for Space Studies three-dimensional model. J. Geophys. Res., 93(D8), 9341-9364.

Hellmer, H. & D. Olbers (1989) A two-dimensional model of the thermohaline circulation under an ice shelf. Antarctic Science, 1(4), 325-336.

Hoffman, J.S., D. Keyes & J.G. Titus (1983) Projecting future sea level rise. Office of Policy and Resource Management, Environmental Protection Agency, Washington, EPA 230-09-007, 121 pp.

Hughes, T. (1973) Is the West Antarctic ice sheet disintegrating? J. Geophys. Res., 78(33), 7884-7910.

Hult, J.L. & N.C. Ostrander (1973) Antarctic icebergs as a global freshwater resource. Rand Rept. R-1255-NSF, 83 pp.

Intergovernmental Panel on Climate Change (1990) Scientific Assessment of Climate Change. WMO/UNEP, J.T. Houghton, G.L. Jenkins and J.J. Ephraums (Eds), Camb. Univ. Press, 26 pp.

Jacobs, S., A.L. Gordon & J.L. Ardai Jr. (1979) Circulation and melting beneath the Ross Ice Shelf. Science, 203, 439-442.

Jacobs, S. (1985) Oceanographic evidence for land ice / ocean interactions in the Southern Ocean. In: Glaciers, Ice Sheets and Sea Level: Effects of a CO_2-Induced Climatic Change, Rep. of Workshop, Seattle, September 13-15, 1984, Nat. Res. Coun., Nat. Acad. Press, Wash. D.C., Dept. of Energy # DOE/ER/60235-1, 116-128.

Jacobs, S.S., R.G. Fairbanks, & Y. Horibe (1985) Origin and evolution of water masses near the Antarctic continental margin: Evidence from $H_2{}^{18}O/H_2{}^{16}O$ ratios in seawater, In: Oceanology of the Ant. Cont. Shelf, Ant. Res. Ser., 43, Am. Geophys. Un., Wash., D.C., 59-85.

Jacobs, S. (1986) Injecting ice-shelf water and air into the deep Antarctic oceans, Nature, 321, 6067, 196-197.

Jacobs, S.S., D.R. MacAyeal & J.L. Ardai, Jr. (1986) The recent advance of the Ross Ice Shelf, Antarctica. J. Glaciol., 32(112), 464-474.

Jacobs, S.S., (1989) Marine controls on modern sedimentation on the Antarctic continental shelf, Mar. Geol., 85, 121-153.

Jacobs, S.S. & J. Comiso (1989) Sea Ice and Oceanic Processes on the Ross Sea Continental Shelf, J. Geophys. Res., 94(C12), 18195-18211.

Jacobs, S.S., W. Haines, J.L. Ardai, Jr., & P. Mele (1989) Ross Sea Oceanographic Data, 1983-1987. Tech. Rep. LDGO 89-5, Lamont-Doherty Geol. Obs. of Columbia University, 482 pp.

Jenkins, A. & C.S.M. Doake (1991) Ice-ocean interaction on Ronne Ice Shelf, Antarctica. J. Geophys. Res., 96(C1), 791-813.

Keys, H.J.R., S.S. Jacobs & D. Barnett (1990) The calving and drift of iceberg B-9 in the Ross Sea, Antarctica. Ant. Sci., 2(3), 243-257.

Lange, M. & H. Kohnen (1985) Ice front fluctuations in the eastern and southern Weddell Sea. Ann. Glaciol., 6, 187-191.

Lewis, E.L. & R. Perkin (1986) Ice pumps and their rates. Jour. Geophys, Res, 91(C10), 11756-11762.

Lingle, C.S. (1985) A model of a polar ice stream and future sea-level rise due to possible drastic retreat of the West Antarctic ice sheet. In: NRC/DOE (1985) Glaciers, ice sheets, and sea level: Effect of a CO_2-induced climatic change, Report of a workshop, Seattle, 13-15 Sept. 1984, Nat'l. Res. Council, Polar Res. Bd., Dept. of Energy (DOE/EV/60235-1), 317-330.

Lingle, C.S., D.H. Schilling, J.L. Fastook, W.S.B. Paterson, & T.J. Brown (1990) A flow band model of the Ross Ice Shelf, Antarctica: Response to CO_2-induced climatic warming. J. Geophys. Res., in press.

MacAyeal, D.R. (1984a) Thermohaline circulation below the Ross Ice Shelf: A consequence of tidally induced vertical mixing and basal melting. J. Geophys. Res., 89(C1), 607-615.

MacAyeal, D.R. (1984b) Potential effect of CO_2 warming on sub-ice-shelf circulation and basal melting. In: Environment of W. Antarctica: potential CO_2-induced changes; report of a workshop, July 1983, Washington, DC, 212-221.

MacAyeal, D.R. (1985) Evolution of tidally triggered meltwater plumes below ice shelves. In: Oceanology of the Antarctic Continental Shelf, Ant. Res. Ser., Vol. 43, Am. Geophys. Un., Wash. DC, 133-143.

Manabe, S., K. Bryan & M.J. Spelman (1990) Transient response of a global ocean-atmosphere model to a doubling of atmospheric carbon dioxide. J. Phys. Oceanogr., 20, 722-749.

Markov, K.K., V.I. Bardin, V.L. Lebedev, A.I. Orlov & I.A. Suyetova (1968) Geografiya Antarktidy [The Geography of Antarctica]. Moskva, Izdatel'stvo Mysl'. [English translation: Jerusalem, Israel Program for Scientific Translations, 1970]

Meier, M.F. (1984) Contribution of small glaciers to global sea level, Science, 226, 1418-1421.

Meier, M.F. (1990a) Role of Land Ice in Present and Future Sea-Level Change. In: Sea-Level Change, Studies in Geophysics, Nat'l Res. Council, Nat'l Acad. Press, Wash., DC, 171-184.

Meier, M.F. (1990b) Reduced rise in sea level. Nature, 343, 115-116.

Michel, R.L., T.W. Linick & P.M. Williams (1979) Tritium and carbon-14 distributions in seawater from under the Ross Ice Shelf Project ice hole. Science, 203(4379), 445-446.

Morgan, V.I. (1972) Oxygen isotope evidence for bottom freezing on the Amery Ice Shelf. Nature, 238, 393-394.

Morgan, V.I. & W.F. Budd (1978) The distribution, movement and melt rates of Antarctic icebergs. In: Iceberg Utilization, A. Husseiny, ed., Proc. of the 1st Int'l Conf., Ames, Iowa, 220-228.

Neal, C.S. (1979) The dynamics of the Ross Ice Shelf revealed by radio echo-sounding. J. Glaciol. 24(90), 295-307.

Neshyba, S. & E.G. Josberger (1980) On the estimation of Antarctic iceberg melt rate. J. Phys. Oceanogr., 10(10), 1681-1685.

NPOC (1987-88) Antarctic Ice Charts, 1987-1988. Naval Polar Oceanography Center, Suitland, MD, 54 pp.

NRC/DOE (1985) Glaciers, ice sheets, and sea level: Effect of a CO_2-induced climatic change, Report of a workshop, Seattle, 13-15 Sept. 1984, Nat'l. Res. Council, Polar Res. Bd., Dept. of Energy (DOE/EV/60235-1), 330 pp.

Orheim, O. (1985) Iceberg discharge and the mass balance of Antarctica. In: NRC/DOE (1985) Glaciers, ice sheets, and sea level: Effect of a CO_2-induced climatic change, Report of a workshop, Seattle, 13-15 Sept. 1984, Nat'l. Res. Council, Polar Res. Bd., Dept. of Energy (DOE/EV/60235-1), 210-215.

Orheim, O. (1990) Extracting climatic information from observations of icebergs in the Southern Ocean. Ann. of Glac., 14, 352 (abstract).

Parkinson, C.L. & R.A. Bindschadler (1984) Response of Antarctic sea ice to uniform atmospheric temperature increases. In: J.E. Hansen and T. Takahashi [Eds], Climate Processes and Climate Sensitivity, Geophys. Mono. 29, Am. Geophys. Un., Wash., DC, 254-264.

Peltier, W.R. & A.M. Tushingham (1989) Global sea level rise and the greenhouse effect: Might they be connected? Science, 244, 806-810.

Pillsbury, R.D. & S.S. Jacobs (1985) Preliminary observations from long-term current meter moorings near the Ross Ice Shelf, Antarctica In: Oceanology of the Antarctic Continental Shelf. Ant. Res. Ser., 43, Washington, DC, Am. Geophys. Un., 87-107.

Pillsbury, R.D., J. Simkin, D. Root & J. Bottero (1989) A compilation of observations from moored current meters, Vol. 16, Current, temperature, conductivity and pressure in the Ross Sea, February 1983 - February 1987. Oregon State Univ., Corvallis, unpublished.

Potter, J.R. & J.G. Paren (1985) Interaction between ice shelf and ocean in George VI Sound, Antarctica. In: Oceanology of the Antarctic Continental Shelf, Ant. Res. Ser., 43, Am. Geophys. Un., Wash. DC, 35-58.

Potter, J.R., M.H. Talbot & J.G. Paren (1988) Oceanic regimes of the ice fronts of George VI Sound, Antarctic Peninsula. Cont. Shelf Res., 8(4), 347-362.

Pozdeyev, V.S. & R.G. Kurinin (1987) Noviye danniye o morfologii ledovoy tolshi i reljefe podliodnogo lozha i morskogo dna v juzhnoy chasti basseina moriya Ueddella (Zapadnaya Antarctida). [New data on the morphology of the ice cover and relief of the subglacial bed and sea bottom in southern part of the Weddell Sea basin (West Antarctica)]. Antarktika, Doklady Komissii, 26, 66-71.

Reeh, N. (1985) Greenland ice sheet mass balance and sea level change. In: NRC/DOE (1985) Glaciers, ice sheets, and sea level: Effect of a CO_2-induced climatic change, Report of a workshop, Seattle, 13-15 Sept. 1984, Nat'l. Res. Council, Polar Res. Bd., Dept. of Energy (DOE/EV/60235-1), 155-171.

Robin, G. (1979) Formation, flow and disintegration of ice shelves. Jour. of Glaciol., 24(90), 259-271.

Robin, G., C.S.M. Doake, H. Kohnen, R.D. Crabtree, S.R. Jordan & D. Moller (1983) Regime of the Filchner-Ronne Ice Shelves, Antarctica. Nature, 302(5909), 582-586.

Robin, G. (1987) Changing the sea level. Ch. 7 in: The Greenhouse Effect, Climatic Change and Ecosystems, B. Bolin, B. Doos, J. Jager & R. Warrick, eds, Scope 29, ICSU, WMO, Wiley, NY, 323-359.

Scheduikat, M. & D. Olbers (1990) A one-dimensional mixed-layer model beneath the Ross Ice Shelf with tidally induced vertical mixing. Antarctic Science 2(1), 29-42.

Schlesinger, M.E. (1985) A numerical simulation of CO_2-induced transient climate change with a coupled atmosphere-ocean general circulation model. In: NRC/DOE (1985) Glaciers, ice sheets, and sea level: Effect of a CO_2-induced climatic change, Report of a workshop, Seattle, 13-15 Sept. 1984, Nat'l. Res. Council, Polar Res. Bd., Dept. of Energy (DOE/EV/60235-1), 267-274.

Schlosser, P. (1986) Helium: A new tracer in Antarctic oceanography. Nature, 321, 233-235.

Schlosser, P., R. Bayer, A. Foldvik, T. Gammelsrod, G. Rohardt & K.O. Munnich (1990) Oxygen 18 and helium as tracers of Ice Shelf Water and water/ice interaction in the Weddell Sea. J. Geophys. Res., 3253-3263.

Shabtaie, S. & C.R. Bentley (1987) West Antarctic Ice Streams draining into the Ross Ice Shelf: Configuration and mass balance. J. Geophys. Res., 92(B2), 1311-1336.

Smith, N.R., D. Zhaoquian, K.R. Kerry & S. Wright (1984) Water masses and circulation in the region of Prydz Bay, Antarctica. Deep-Sea Res., 31(9), 1121-1147.

Stouffer, R., S. Manabe & K. Bryan (1989) Interhemispheric asymmetry in climate response to a gradual increase of atmospheric CO_2. Nature, 342, 660-662.

St.Pierre, D.B. (1989) On the effectiveness of the production of Antarctic bottom water in the Weddell and Ross Seas. Master's Thesis, Naval Postgraduate School, Monterey, 125 pp.

Thomas, R.H. & P.H. Coslett (1970) Bottom melting of ice shelves and the mass balance of Antarctica. Nature, 228(5266), 47-49.

Thomas, R.H. & D.R. MacAyeal (1982) Derived characteristics of the Ross Ice Shelf, Antarctica. J. Glaciol., 28(100), 397-412.

Thomas, R.H., T.V. Martin & H.J. Zwally (1983) Mapping ice-sheet margins from radar altimetry data. Ann. of Glac., 4, 283-288.

Thomas, R.H. (1987) Future sea-level rise and its early detection by satellite remote sensing. Prog. Oceanog., 18, 23-40.

Trumbore, S., S.S. Jacobs and W.M. Smethie, Jr. (1991) Chlorofluorocarbon evidence for rapid ventilation of the Ross Sea. Deep-Sea Res., in press.

Visser, A.W., & S.S. Jacobs (1987) Annual mean estimates of heat transport into the Ross Sea sub-ice cavity. IUGG/IAPSO Symposium, Vancouver, Aug. 1987 (abstract).

Wadhams, P. (1988) Winter observations of iceberg frequencies and sizes in the South Atlantic Ocean. J. Geophys. Res., 93(C4), 3583-3590.

Weeks, W.F. & M. Mellor (1978) Some elements of iceberg technology. In: Iceberg Utilization, A. Husseiny, ed., Proc. of the 1st Int'l Conf., Ames, Iowa, pp. 45-98.

Weiss, R.F., H.G. Ostlund, & H. Craig (1979) Geochemical studies of the Weddell Sea. Deep-Sea Res., 26(10A), 1093-1120.

Wexler, H. (1960) Heating and melting of floating ice shelves. J. Glaciol., 3, 626-645.

Zakharov, V.G. (1988) Fluctuations in ice shelves and outlet glaciers in Antarctica. Polar Geog. and Geol., 12, 297-311. [Translated from Materialy glyatsiologicheskikh issledovaniy, 62, pp. 41-51 (1988)].

Zotikov, I.A. (1963) Bottom melting in the central zone of the ice shield on the Antarctic continent and its influence upon the present balance of the ice mass. Bull. Int'l Assoc. Scient. Hydrol., 8e Annee, No. 1, p. 36-44.

Zotikov, I.A., V.S. Zagorodnov & J.V. Raikovsky (1980) Core drilling through the Ross Ice Shelf (Antarctica) confirmed basal freezing. Science, 207, 1463-1465.

Zwally, H.J., S. Stephenson, R.A. Bindschadler & R.H. Thomas (1987) Antarctic ice shelf boundaries and elevations from satellite radar altimetry. Ann. of Glac., 9

Zwally, H.J. (1989) Growth of the Greenland ice sheet: Interpretation. Science, 246, 1589-1591.

Feature	Area (km^2)	Rate (cm yr^{-1})	Volume (km^3 yr^{-1})
Filchner-Ronne I. S.	403,000	45	181
Ross Ice Shelf	446,000	19	85
George VI Ice Shelf	25,000	210	53
Amery Ice Shelf	19,000	-60	-11
seaward 100km, all ice shelves	650,000	35	227
ice shelf fronts	4,500	350	16
runoff	280,000	13	36
grounded ice sheet	12,000,000	0.35	42
			Total: 629

Table 1. A preliminary estimate of melting around and beneath the Antarctic ice sheet. The seaward 100 km includes the outer 100 km of the Filchner-Ronne, Ross and Amery Ice Shelves. See text for melt rate rationale and literature sources.

Figure Captions

Fig. 1. Annual volume of large icebergs (>28 km) calved from the Antarctic ice sheet, 1979-1989. From NOAA and DMSP satellite data interpreted by the Navy-NOAA Joint Ice Center (e.g., NPOC, 1987-1988). Most iceberg volumes were calculated by assuming an elliptical area and a thickness of 250 m. The other bars show the 1979-1989 large berg average (~1295 km^3 yr^{-1}), and the ice equivalent of 1.5 mm yr^{-1} global sea level change. Accumulation on the ice sheet [1962.7 x 10^{12} kg, from Giovinetto and Bentley (1985) plus 226 km^3 yr^{-1} for the Antarctic Peninsula from Doake, 1985] is indicated by a dashed line at 2535 km^3 yr^{-1}.

Fig. 2. Annual mean current velocity (top panel) and ocean heat transport (bottom panel), derived from sixteen Aanderaa current meters on nine instrument arrays (A-H) bottom-moored along the Ross Ice Shelf front from February 1984 through January, 1985. Negative contours are shaded and indicate flow into the sub-ice shelf cavity. Data from Pillsbury et al. (1989) and Visser and Jacobs (1987).

Fig. 3. Distribution of melting and accumulation rates at the Filchner Ice Shelf base, from a 2-D thermohaline ocean circulation model by Hellmer and Olbers (1989). The standard simulation, for high salinity shelf water at -1.92°C and $34.72\,^{o}/_{oo}$, is shown by the heavy line. Dashed lines illustrate the altered basal mass fluxes when the shelf water is colder and saltier (circles) or warmer and fresher (triangles), in each case by .02°C and $.02\,^{o}/_{oo}$. The lettered bars refer to estimates by previous investigators, some for other ice shelves. A steady-state model of the basal mass flux from glaciological measurements along a 760 km Ronne Ice Shelf flowline shows a similar distribution of melting and freezing areas (fig. 10 in Jenkins and Doake, 1990).

Fig. 4 Ice Shelf Water (ISW) outflow from beneath the Ross Ice Shelf overrides the High Salinity Shelf Water (HSSW) formed by surface freezing in the open Ross Sea. The ISW envelope (heavy dashed line) is defined here by the portions of February 1984 vertical profiles along the ice shelf front where the temperature was below the sea surface freezing point. Data from Jacobs et al. (1989). Small circles denote local temperature minima during several austral summers, with the solid circles (1984 data) joined by light lines. Large squares indicate the locations of long-term (7-12 month) current measurements, from Pillsbury et al. (1989).

Fig. 5 A simplified representation of sea level change over the past century, and various projections for the next century. In the original sources, most records and predictions are not linear and may have been given as a single value. The 35 cm rise by 2050 of Meier 1990(b) incorporates current and future Antarctic ice sheet growth equivalent to a sea level fall of 0.75 and 5.0 mm yr^{-1}, respectively. The short dashed lines approximate present-day sea level rise (1.5 mm yr^{-1} or 635 km^3 yr^{-1} ice equivalent from Fig. 1), and the largest meltwater pulse from the latest deglaciation, spread over a period of 10^3 yr (24 mm yr^{-1}, derived from Fairbanks, 1989).

Fig. 6 Schematic depiction of a possible SeaRISE transect through the Ross Ice Shelf and across the continental shelf and slope. The vertical lines are drawn to indicate that ocean station profiles and access holes through the ice shelf above the ISW plume should be concentrated near the grounding lines, ice rises, ice front and the continental shelf break. Full sampling for transient tracer and stable isotope geochemistry is essential both across and along the ice front. Beneath the ocean surface, time series recordings of currents, temperatures, and sea ice thickness could be recovered by ship; thermistor chains and other devices in and below the shelf ice could transmit data by satellite. It would be desirable for the ice shelf transect to branch onto both the West and East Antarctic ice sheets, where different water depths and circulation patterns are likely near the grounding lines.

BLANK
PAGE

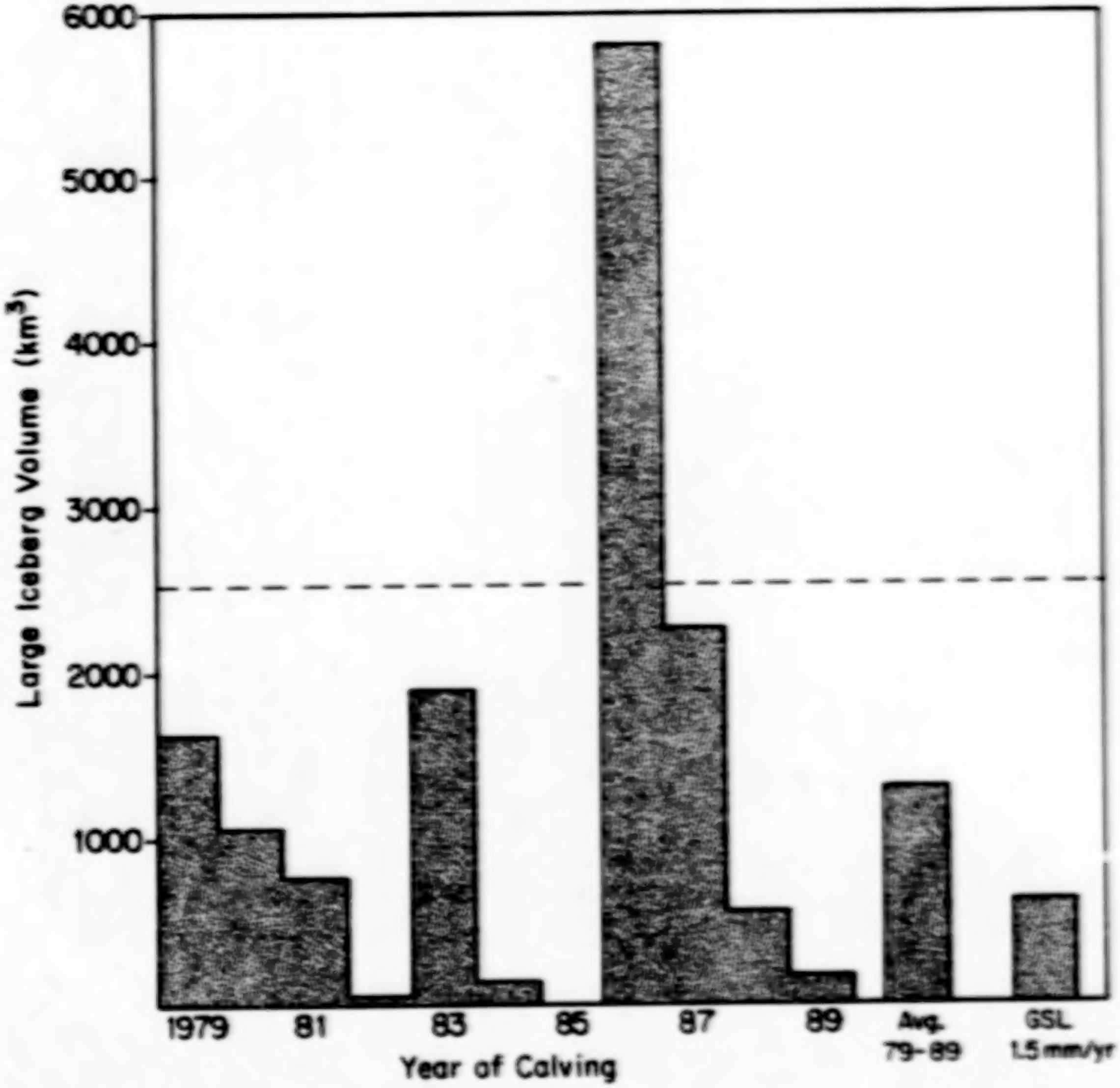

6000
5000
4000
3000
2000
1000
Large Iceberg Volume (km³)
1979
81
83
85
87
89
Avg. 79-89
GSL 1.5 mm/yr
Year of Calving

Fig 1

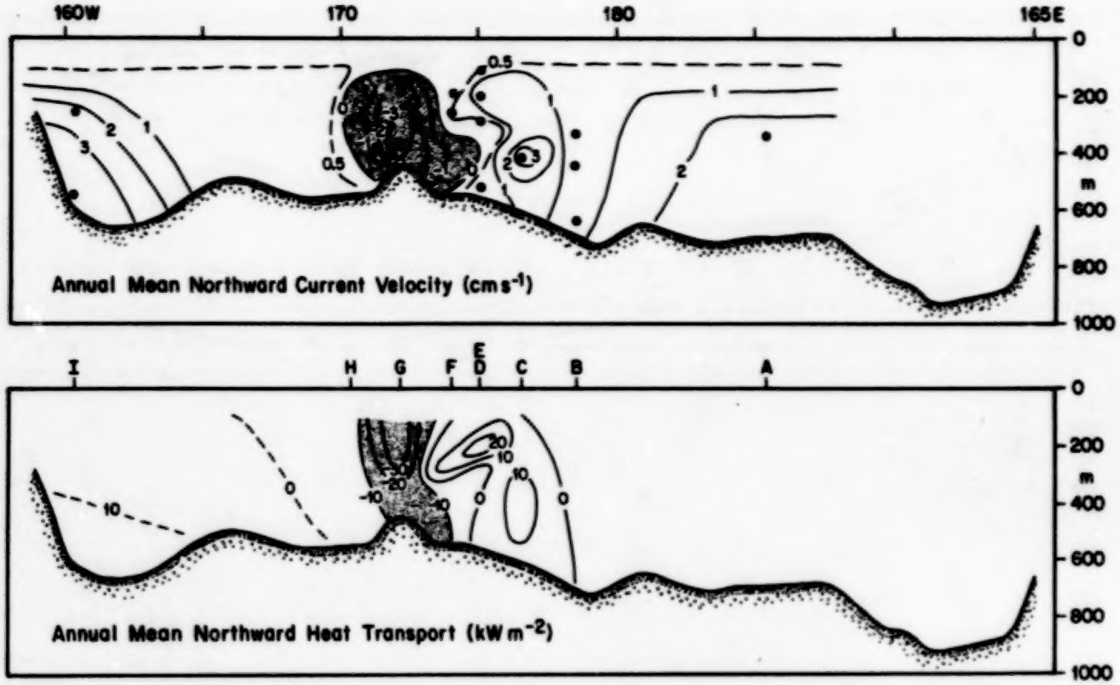

160W
170
180
165E
0
200
400
m
600
800
1000
0.5
1
2
3
Annual Mean Northward Current Velocity (cm s-1)
I
H
G
F
E
D
C
B
A
10
0
-10
20
-30
Annual Mean Northward Heat Transport (kW m-2)

Fig 2

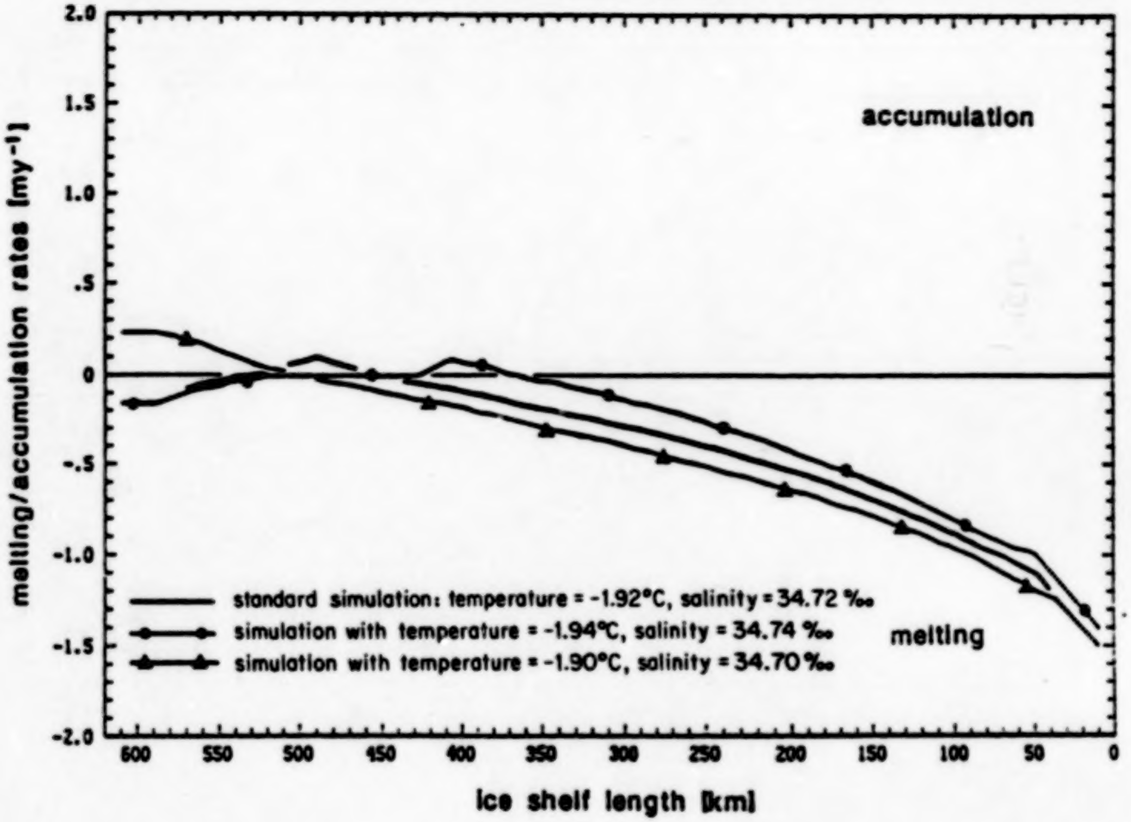

2.0
1.5
1.0
.5
0
-.5
-1.0
-1.5
-2.0
melting/accumulation rates [my^{-1}]
accumulation
melting
standard simulation: temperature = -1.92°C, salinity = 34.72 ‰
simulation with temperature = -1.94°C, salinity = 34.74 ‰
simulation with temperature = -1.90°C, salinity = 34.70 ‰
600
550
500
450
400
350
300
250
200
150
100
50
0
ice shelf length [km]

Fig 3

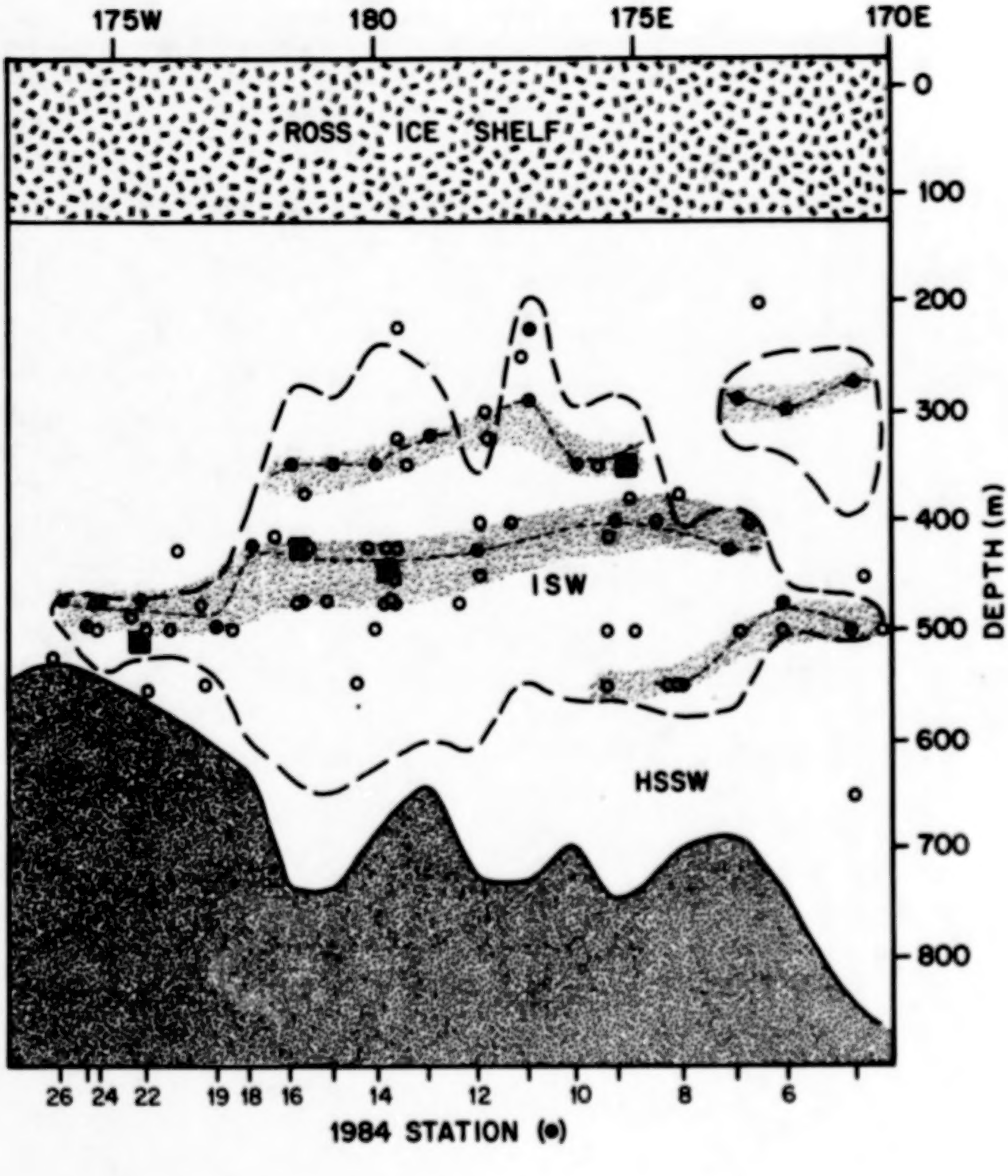

175W
180
175E
170E
ROSS ICE SHELF
0
100
200
300
400
500
600
700
800
DEPTH (m)
ISW
HSSW
26
24
22
19
18
16
14
12
10
8
6
1984 STATION (●)

Fig 4

BLANK
PAGE

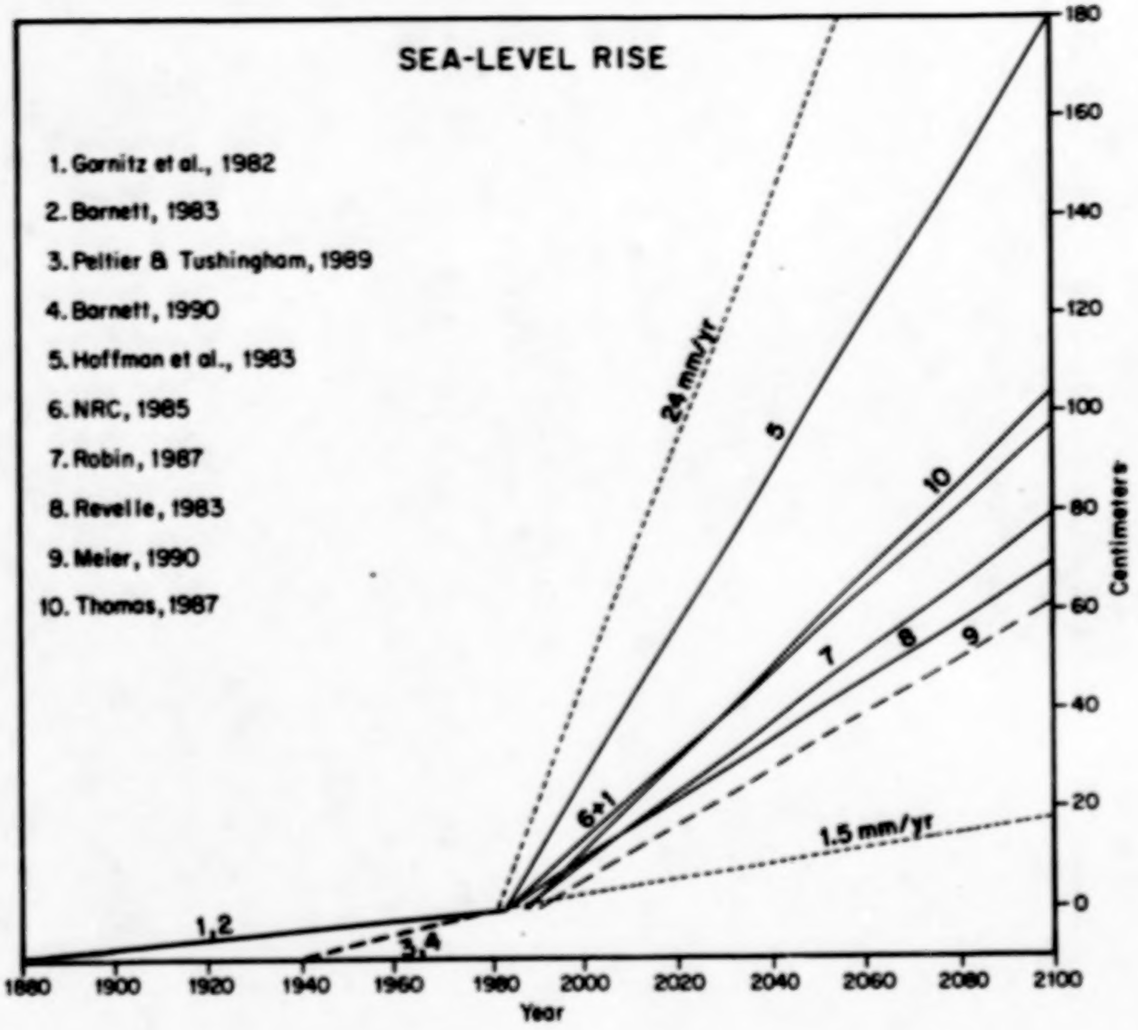

SEA-LEVEL RISE
1. Gornitz et al., 1982
2. Barnett, 1983
3. Peltier & Tushingham, 1989
4. Barnett, 1990
5. Hoffman et al., 1983
6. NRC, 1985
7. Robin, 1987
8. Revelle, 1983
9. Meier, 1990
10. Thomas, 1987
24 mm/yr
5
10
7
8
9
6+1
1.5 mm/yr
1,2
3,4
180
160
140
120
100
80
60
40
20
0
Centimeters
1880
1900
1920
1940
1960
1980
2000
2020
2040
2060
2080
2100
Year

Fig 5

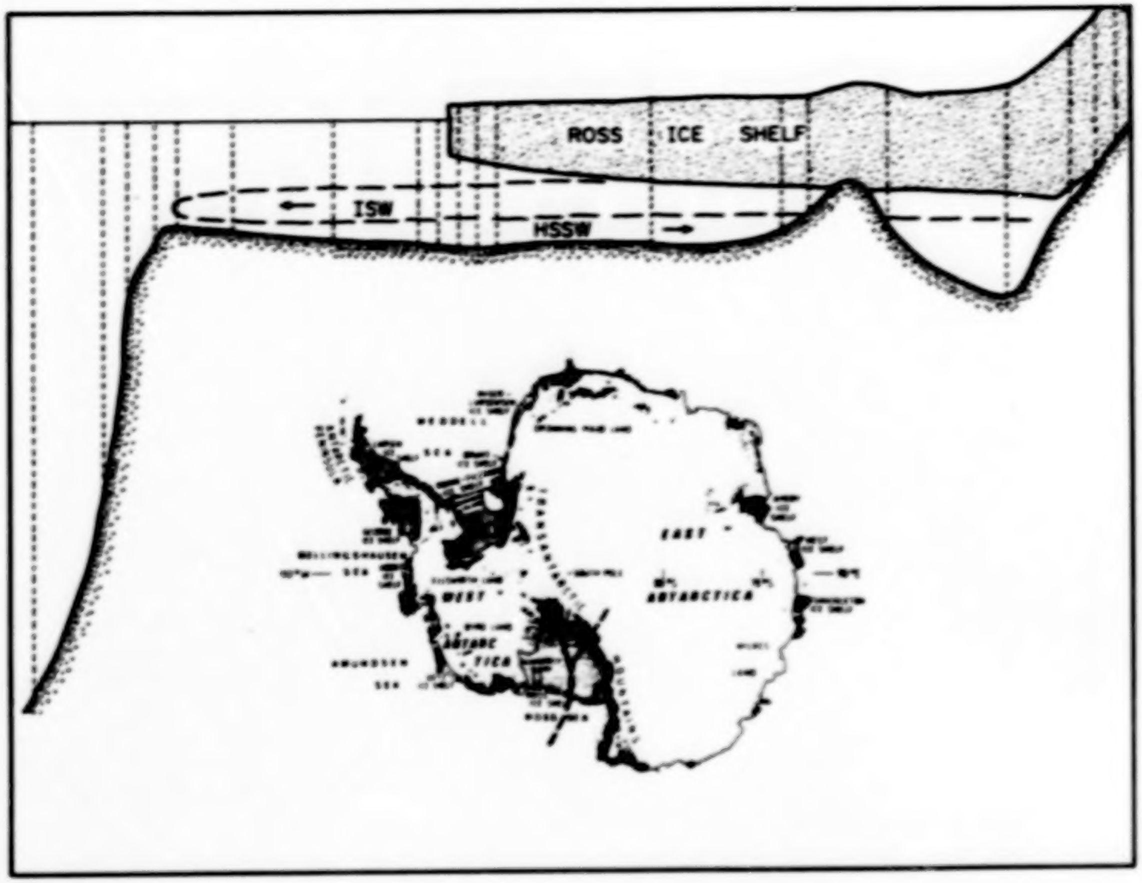

ROSS ICE SHELF
ISW
HSSW
WEDDELL
SEA
BELLINGSHAUSEN
SEA
AMUNDSEN
SEA
WEST
ANTARC
TICA
EAST
ANTARCTICA

Fig 6

BLANK PAGE

Terrestrial Geophysics in the SeaRISE Project

C. R. Bentley
Geophysical and Polar Research Center
University of Wisconsin-Madison

Radar Sounding

Radar sounding serves multiple purposes. The most general and obvious is mapping ice thickness and the surface and bedrock topography of the ice sheet. Determining the surface and basal elevations of the ice sheet requires determining the height of the sounding aircraft, although measuring ice thickness does not. In addition to this basic mapping, there are many problem-oriented uses of radar sounding. Specific applications to the problems of the grounded and floating parts of the West Antarctic ice sheet are as follows.

1) Examination of the variations in bottom echo strength as a means of distinguishing between wet and frozen beds and perhaps locating ponded water beneath the inland ice.
2) Mapping ice rises, ice rumples, and ice rafts on the ice shelves. Ice rises and ice rafts are characterized by a lack of "clutter", that is, scattering on the radar return from crevasses buried beneath the surface. Ice rumples do have clutter, as do the ice streams and portions of the ice shelf that have arisen from the ice streams.
3) Mapping the boundaries of ice streams, and zones in the inland ice that are incipient ice streams, by the location of clutter zones. The work to date has shown very complex patterns of cluttered and clutter-free ice around the heads of the West Antarctic ice streams.
4) Mapping of "debris tracks", and identification of their sources. "Debris

tracks" are internal reflectors arising from either rock debris or bottom crevasses that can be associated with a particular geographical location as a source, and then traced downstream. The divergence between debris tracks and the present day flow lines acts as a record of changing flow patterns.

5) Measurements, with short pulse radar systems on the ground, of the depth to buried crevasses as a means of determining how long it has been since those crevasses were open at the surface. In the case of ice stream C, such measurements have provided an estimate of how long ago it was that ice stream C ceased to be active.

6) Profiling of internal layers for analysis in terms of changing configuration of the ice sheet, as was done for the vicinity of Byrd Station.

7) Local radar mapping in association with strain and gravity measurements. In the case of strain measurements, thickness variations need to be known to provide adequate interpretation of the strain rates. In the case of gravity measurements, radar surveys are needed to provide the subglacial topographic corrections necessary for the determination of regional characteristics.

8) Measurement of anisotropy in the ice through its effect on radio wave polarizations. Particular attention in this regard should be paid to the depolarization of internal reflections that can yield clues to variation of crystal fabric with depth.

9) Mapping the radar diffraction pattern on a survey grid to be compared with later remapping to see how the diffraction pattern has changed

relative to the surface. Comparison with satellite-determined velocities should reveal whether the diffraction pattern is fixed relative to a stationary bedrock, relative to the base of the ice, or relative to reflection horizons deep within the ice above the bedrock. This is a point of great interest in determining the contributions of bed deformation and strain within the ice and to the total ice movement.

10) Determination of the small-scale statistical roughness of the basal reflecting surfaces can provide useful information about basal sliding. A digital radar recording system makes it possible to determine such statistical characteristics. We hope that when detailed comparisons are made between the characteristics of zones where there is a basal deforming till and zones where there is not, it may become possible to determine the presence of a deforming layer by means of analyses of airborne radar along.

Seismic Shooting

The purpose of seismic shooting, in addition to water depth measurements on floating ice, is to provide information about the internal physical characteristics of the ice sheet, the rock beneath it, and the interface between the two. The point of greatest interest in the current measurements is the extent and nature of a deforming layer immediately below the ice, observed by seismic measurements to be a zone of high porosity and high hydrologic pore pressure. Seismic anisotropy is also of interest because of the relatively large effect of crystal orientation on seismic wave speeds, and because of the importance of crystal orientation in ice flow.

High-resolution seismic reflection profiles utilize both compressional and shear waves to yield the details of the subglacial material and the ice-rock interface. If a subglacial layer is sufficiently thick (more than 5 m), seismic travel times can be used to calculate both its porosity and pore-water pressure. Additionally, studies of the amplitudes of compressional waves that are converted to shear waves upon reflection at the base of the ice may prove useful for detecting a millimeters-thick basal water layer. The comparison of seismic reflection times with radar reflection times at a large number of stations should give a clue to anisotropy in the ice, since seismic velocities are markedly affected by anisotropy whereas electromagnetic wave velocities are not. But the principal way of determining anisotropy is through wide-angle reflection sounding, by means of which the wave velocity can be determined at different angles of incidence and along different azimuths.

Short refraction shooting yields detailed determinations of the wave velocities as a function of depth in the firn layers. Experience has shown that the correlation between wave velocity and density is excellent. The "critical depth of densification" and the depth of the firn-ice boundary can be found by analysis of the wave-velocity gradient as a function of depth. In addition, long-term mean accumulation rates can be estimated from the seismic data, particularly where horizontal strain rates have been determined.

Seismic long-refraction and deeply penetrating reflection shooting, while aimed primarily at the geological objective of determining the upper crustal structure beneath the ice, serve a glaciological purpose as well in yielding information about the stratigraphy, age, and thickness of subglacial sediment, hence about the glacial history of the region. Subglacial layers that are probably composed of Cenozoic marine sediments and whose aggregate thickness may be a kilometer or more are found extensively in West Antarctica.

Passive Seismic Studies

Passive seismic monitoring of microearthquakes can be used to study brittle fracture within the ice or the rock beneath it. Common parameters available from these studies are fault location, orientation, and displacement, as well as the size of the rupture area, stress drop, and energy released. The parameters derived from microearthquakes originating in the chaotic zones that bound the active ice streams should increase our understanding of the stress regime in these regions as well as the contribution of brittle fracture to the observed strain rates. Similarly, parameterization of seismic events arising from the base of the active ice streams may give insight into the erosional processes that are critical to the mass balance of any deforming sediments. In one relatively inactive region at least (ice stream C) faulting may contribute substantially to ice motion.

Electrical Resistivity

There is a large contrast in electrical resistivity between ice or permafrost on the one hand and liquid water or wet rock on the other. Thus, electrical resistivity profiles have the potential capability of revealing the depth to the melting point, whether that melting point is found at the base of the ice or in the subglacial rock.

Extensive studies on the Ross Ice Shelf have suggested that a deep layer with electrical resistivity much higher than in the overlying ice exists at or near the bottom of ice streams and outlet glaciers. If this is so, it is possible that it arises from annealing of the ice due to its strain history. Other increases in electrical resistivity with depth may be associated with the Wisconsin-Holocene Boundary. If the latter association can be confirmed, it will then be possible to estimate the depth to that boundary from resistivity measurements.

Gravity

Gravity anomalies, particularly combined with seismic measurements, are an effective tool for determining deeper crustal structure. Anomalies averaged over extensive areas are useful also for their potential to reveal isostatic imbalance, which is a measure of average glacial change over the last several thousand years. Older studies of this kind in West Antarctica suffered from the large errors in free-air gravity anomalies that stem from poorly known elevations of the surface stations. Satellite determinations of elevation reduce those errors by at least an order of magnitude. In fact, modern navigational techniques will soon make it possible to conduct gravity surveys by small aircraft in West Antarctica.

Late Wisconsin and Early Holocene Glacial History, Inner Ross Embayment, Antarctica

GEORGE H. DENTON

Department of Geological Sciences and Institute for Quaternary Studies, University of Maine, Orono, Maine 04469

JAMES G. BOCKHEIM AND SCOTT C. WILSON

Department of Soil Science, University of Wisconsin, Madison, Wisconsin 53706

AND

MINZE STUIVER

Quaternary Isotopes Laboratory, Quaternary Research Center, University of Washington, Seattle, Washington 98195

Lateral drift sheets of outlet glaciers that pass through the Transantarctic Mountains constrain past changes of the huge Ross ice drainage system of the Antarctic Ice Sheet. Drift stratigraphy suggests correlation of Reedy III (Reedy Glacier), Beardmore (Beardmore Glacier), Britannia (Hatherton/Darwin Glaciers), Ross Sea (McMurdo Sound), and "younger" (Terra Nova Bay) drifts; radiocarbon dates place the outer limits of Ross Sea drift in late Wisconsin time at 24,000–13,000 yr B.P. Outlet-glacier profiles from these drifts constrain late Wisconsin ice-sheet surface elevations. Within these constraints, we give two extreme late Wisconsin reconstructions of the Ross ice drainage system. Both show little elevation change of the polar plateau coincident with extensive ice-shelf grounding along the inner Ross Embayment. However, in the central Ross Embayment one reconstruction shows floating shelf ice, whereas the other shows a grounded ice sheet. Massive late Wisconsin/Holocene recession of grounded ice from the western Ross Embayment, which was underway at 13,040 yr B.P. and completed by 6600–6020 yr B.P., was accompanied by little change in plateau ice levels inland of the Transantarctic Mountains. Sea level and basal melting probably controlled the extent of grounded ice in the Ross Embayment. The interplay between the precipitation (low late Wisconsin and high Holocene values) and the influence of grounding on outlet glaciers (late Wisconsin thickening and late Wisconsin/Holocene thinning, with effects dying out inland) probably controlled minor elevation changes of the polar plateau. © 1989 University of Washington.

INTRODUCTION

The Antarctic Ice Sheet is one of the most prominent physical features on our planet. A knowledge of its past behavior can help solve the cause of late Quaternary ice ages. For example, Antarctic ice-sheet reconstructions form part of global snapshots of the last glacial maximum that are a basis for atmospheric modeling experiments. In the pinpointing of physical mechanisms of paleoclimate changes, the influence of the Antarctic Ice Sheet (including potential fringing shelf ice) on ice-age climates can be determined by prescribing reconstructed ice-sheet geometry as the sole variable input into global climate modeling experiments that compare ice-age and modern climates. This reconstructed geometry can also give the contribution of Antarctic ice to sea-level and marine oxygen isotope records. A knowledge of Antarctic ice-volume changes constrains interpretations of marine oxygen isotope records and, hence, Northern Hemisphere ice-sheet history. Finally, the paleoclimate record from this far-southern region is an important part of the global data bank used to test ice-age theories, particularly with respect to the relative timing of climate changes between polar hemispheres.

To address these issues, we describe here late Quaternary paleoclimate, ice-volume change, and dynamics of the Ross ice drainage system using data from a powerful combination of drift sheets and ice cores. The Ross ice drainage system encompasses about one-fourth of the surface area of the Antarctic Ice Sheet (Fig. 1). Inland ice divides encircling this drainage system are 5700 km long and display numerous domes and saddles. Ice from the high East Antarctic plateau flows seaward in outlet glaciers that pass through the Transantarctic Mountains into the Ross Ice Shelf or Ross Sea. Ice from the marine-based West Antarctic ice sheet flows through major ice streams into the Ross Ice Shelf. Three deep ice cores with dated paleoclimatic records come from within the Ross ice drainage system (Jouzel *et al.*, 1989). Drift sheets occur alongside Transantarctic Mountain outlet glaciers deep within the Ross Embayment.

We start by selecting Transantarctic Mountains outlet glaciers that lie along ice flowlines leading inland to major domes and ice-core sites. From lateral drift sheets we then reconstruct former longitudinal ice-surface profiles of these outlet glaciers. For age control of these profiles, we develop a numerical chronology for several drift sheets in the Taylor Valley/McMurdo Sound and Hatherton/Darwin Glacier areas

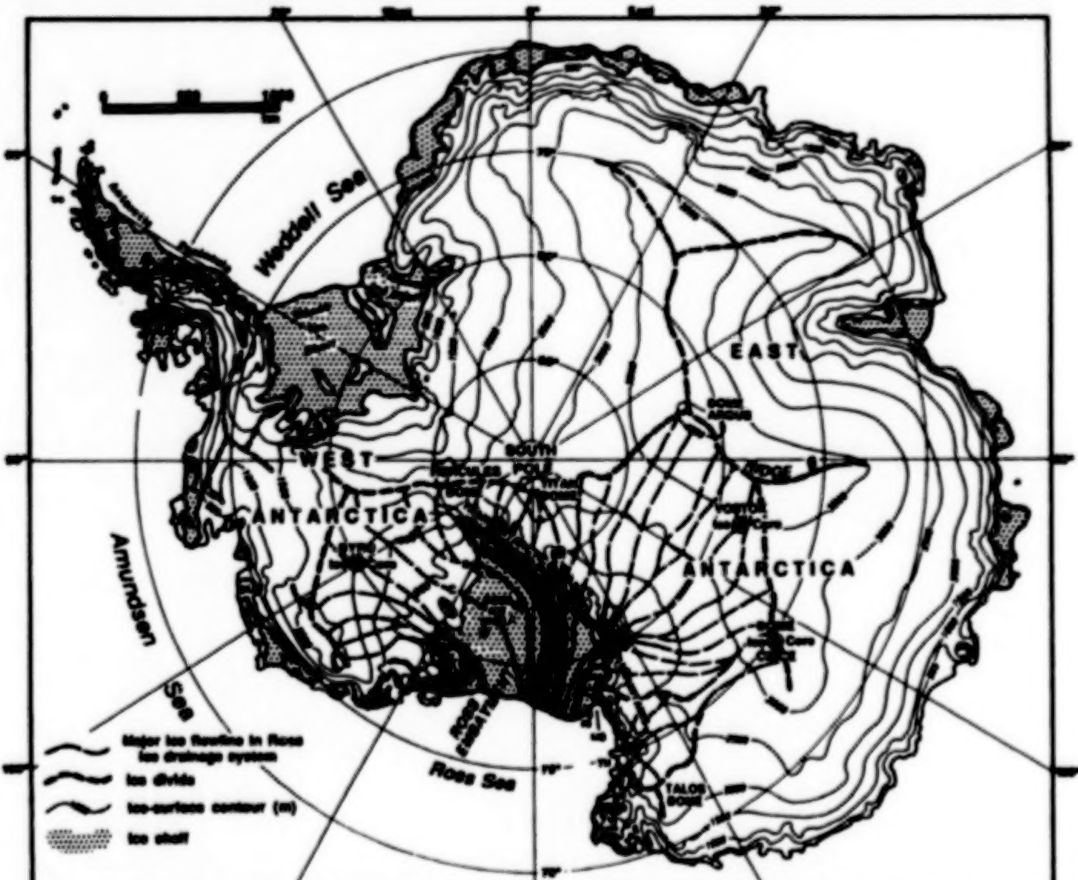


FIG. 1. Index map of Antarctica. Flowlines shown only for the Ross ice drainage system, which is defined as the portion of the Antarctic Ice Sheet that flows into the Ross Embayment. A,B,C,D, and E are West Antarctic ice streams. RG, Reedy Glacier; BG, Beardmore Glacier; DR, Dominion Range ice cap; HG, Hatherton Glacier; DG, Darwin Glacier; DV, Dry Valleys; RI, Ross Island; MS, McMurdo Sound; and TN, Terra Nova Bay. Adapted from Drewry (1983).

(Fig. 1). This numerical chronology is then supplemented by a relative chronology that employs surface boulder weathering and soil development to correlate drift sheets and, hence, ice profiles among the selected outlet glaciers. The late Wisconsin longitudinal profiles of these selected Transantarctic outlet glaciers constrain elevation changes of the inland ice plateau through the last glacial cycle. Hence, they bear directly on Antarctic ice-volume changes as well as on the interpretation of paleotemperatures from interior ice cores.

The outlet glacier profiles also bear on the configuration of grounded and floating ice in the marine portion of the Ross ice drainage system. This problem has been of interest for nearly 150 yr. In 1841 and 1842, Ross (1847) mapped the edge of the Ross Ice Shelf, or the "Great Icy Barrier." Scott (1905, Vol. 2, pp. 422–425) postulated that "when the Southern glaciation was at a maximum . . . the Great Barrier was a very different formation from what it is today . . . the huge glacier, no longer able to float on a sea of 400 fathoms, spread out over the Ross Sea, completely filling it with an immense sheet of ice." Scott (1905, Vol. 2, p. 425) further postulated that, during recession, the ice sheet became buoyant and broke away gradually so that the " . . . Barrier is the remains of the great ice-sheet." David and Priestley (1914, Fig. 46, Pl. XCV) also postulated northward expansion so that at the time of maximum glaciation the surface of the grounded Ross Ice Barrier (Ross Ice Shelf) reached heights of 1000 ft (305 m) in McMurdo Sound and filled the Ross Sea with a grounded "great ice sheet" for at least 320 km north of the current Barrier (Shelf) edge. In sharp contrast, Debenham (1921, Fig. 11) interpreted glacial deposits in the McMurdo Sound area in terms of local glaciation. In recent years this difference of opinion concerning extensive (Denton and Armstrong, 1968; Denton *et al.*, 1970, 1971, 1975; Denton and Borns, 1974; Mercer, 1968, 1972; Kellogg *et al.*, 1979) as opposed to restricted (Péwé, 1960; Fillon, 1975) grounded ice in the Ross Sea has resulted in differing reconstructions of late Wisconsin ice extent in the Ross Embayment. The CLIMAP maximum reconstruction is based largely on geological data and shows widespread grounded ice in the Ross Embayment (Stuiver *et al.*, 1981, pp. 376–380). The CLIMAP minimum reconstruction (Stuiver *et al.*, 1981, p. 375) shows no change in the Ross Embayment and is based on glaciological data that imply little change in the West Antarctic Ice Sheet during the last glacial cycle (Whillans, 1976). In an attempt to reconcile the geologic and glaciologic data, Drewry (1979) reconstructed a restricted complex of local domes and a large floating ice shelf that extends deep into the Ross Embayment to the Transantarctic Mountains at the mouth of major East Antarctic outlet glaciers. Our new ice-surface profiles from the Transantarctic Mountains show deficiencies in all these existing reconstructions. As a result we give here new reconstructions of the minimum and maximum extent of grounded ice in the Ross Embayment permitted by our glacial geologic data in the adjacent Transantarctic Mountains.

LATE WISCONSIN SUBSTAGE

Ice-Surface Profiles

Figures 2, 3, and 4 show former longitudinal profiles of late Quaternary age for selected outlet glaciers of the East Antarctic Ice Sheet that flow through the Transantarctic Mountains to the Ross Ice Shelf. These selected outlet glaciers are among the few in the Transantarctic Mountains that have late Quaternary lateral drift sheets. Reedy Glacier lies along a flowline that leads from the Ross Ice Shelf inland through Ice Stream A to the interior Hercules Dome (Fig. 1). Beardmore Glacier is on a flowline that projects inland to the Titan Dome. The Hatherton Glacier is tributary to the Darwin Glacier, which drains ice from Dome Circe.

Former longitudinal profiles were derived from the outer limits of drift sheets correlated among ice-free areas adjacent to

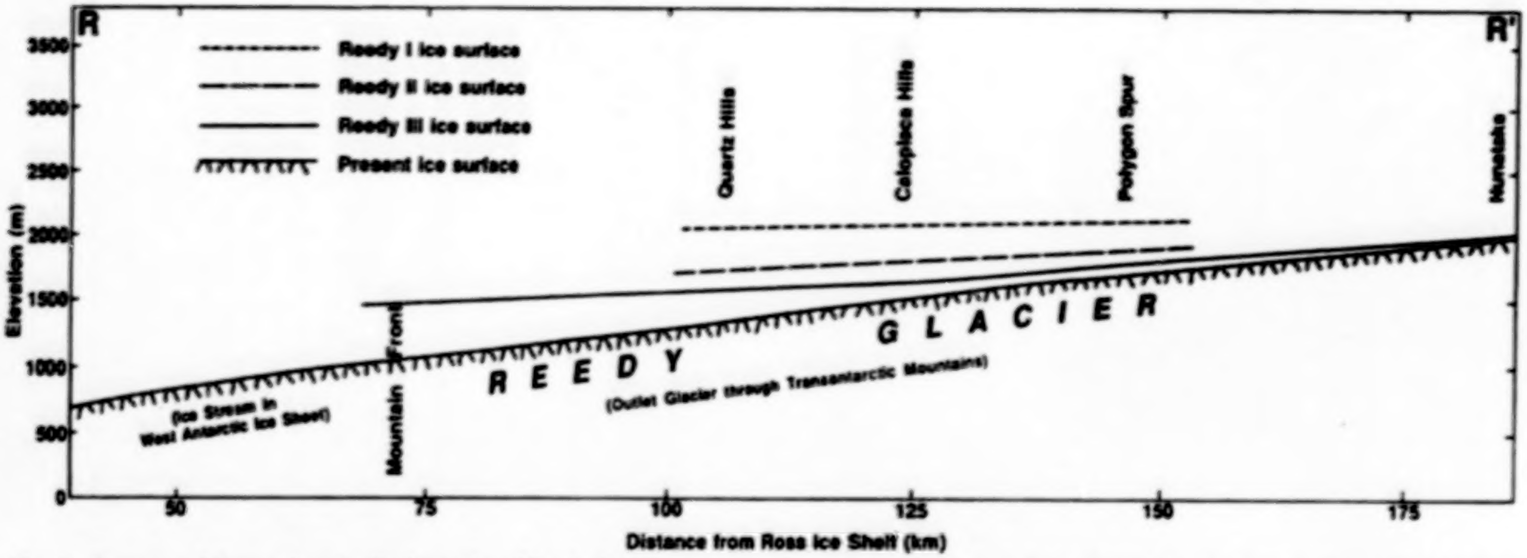


FIG. 2. Present and former longitudinal profiles of Reedy Glacier. Position of profile is shown as R–R' in Figure 12a. Profiles are from data in Mercer (1968).

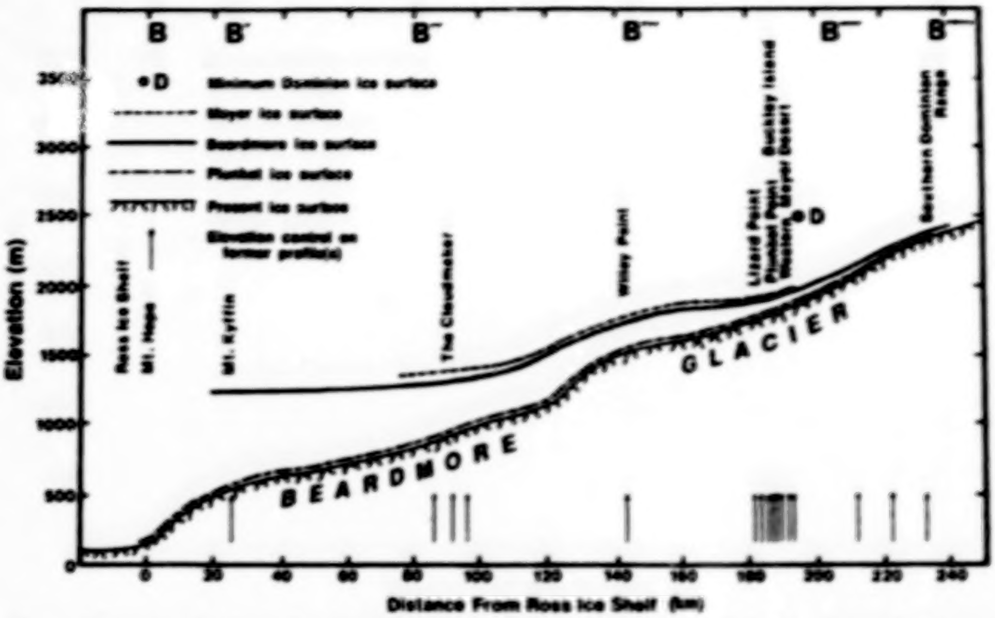


FIG. 3. Present and former longitudinal profiles of Beardmore Glacier. Position of profiles shown as B–B′′′′′ in Figure 12a. See Denton *et al.* (1989) for details.

these outlet glaciers. Details for the Beardmore and Hatherton/Darwin Glacier areas are reported by Denton *et al.* (1989) and Bockheim *et al.* (1989). Drift sheets in these two areas were differentiated by surface morphology, surface boulder weathering, and soil development (staining and cohesion depths, solum thickness, morphogenetic salt stage, and weathering stage). We reconstructed former profiles of Reedy Glacier from data in Mercer (1968). We are confident that these former profiles of Reedy Glacier are directly comparable to our profiles of Beardmore and Hatherton Glaciers, because our mapping of drift units at Beardmore Glacier (Denton *et al.*, 1989) afforded results very similar to those obtained in the same field area earlier by Mercer (1972).

Critical to our discussion are the Reedy II and III profiles of Reedy Glacier (Fig. 2) (Mercer, 1968); the Beardmore and Meyer profiles of Beardmore Glacier (Fig. 3) (Denton *et al.*, 1989); and the Britannia I, Britannia II, and Danum profiles of Hatherton Glacier (Fig. 4) (Bockheim *et al.*, 1989). All of these profiles show only slight thickening at glacier heads near the polar plateau of the East Antarctic Ice Sheet, contemporaneous with substantial thickening at glacier mouths near the Ross Ice Shelf. Our leading interpretation of these ice profiles (Denton *et al.*, 1989; Bockheim *et al.*, 1989) is that they reflect ice-shelf grounding in the southwestern and southern Ross Embayment. An alternative explanation of increased ice flow from higher precipitation on the polar plateau is less plausible. The maximum potential elevation increase along flowlines on the polar plateau (Fig. 1) would be 30 m inland of Reedy Glacier during the Reedy III episode, 35–40 m inland of Beardmore Glacier during the Beardmore episode, and 100 m inland of Hatherton and Darwin Glaciers during the Britannia II episode. From drift geometry, we argued that the Dominion Range ice cap near the polar plateau was retracted when Beardmore Glacier stood at its expanded Beardmore profile (Denton *et al.*, 1989). A strong inference is that precipitation was lower at that time on the Dominion Range ice cap and the adjacent polar plateau. If this inference is correct, the actual elevation change along the polar plateau flowline inland of Beardmore Glacier would depend on the inter-

BLANK
PAGE

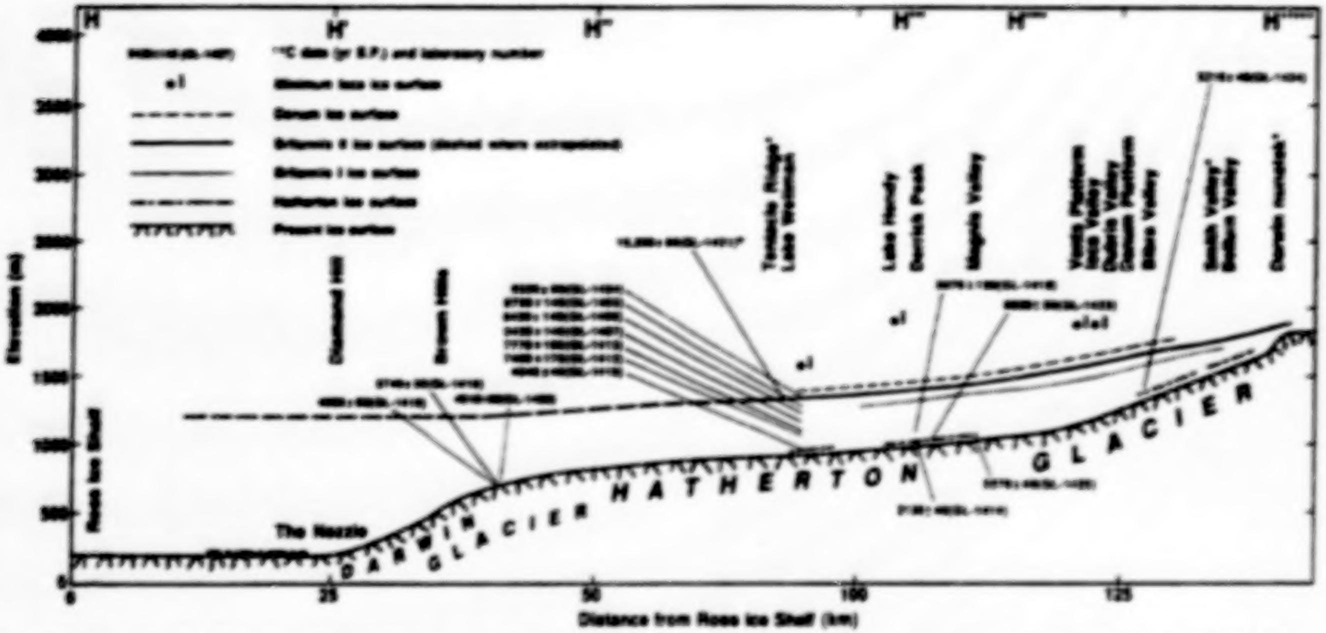


FIG. 4. Present and former longitudinal profiles of Hatherton Glacier. Position of profiles is shown as H–H′′′′′ in Figure 12a. See Bockheim *et al.* (1989) for details.

play between the grounding effect and the lowered precipitation. Decreased precipitation could result from lower air temperatures and/or more-distant open water.

The McMurdo Sound/Dry Valleys area is unique in the Transantarctic Mountains for its extensive ice-free terrain. Only here do East Antarctic outlet glaciers fail to reach the Ross Embayment either today or during late Quaternary glaciations. Rather, East Antarctic ice is blocked by the Royal Society Range or, in the case of Taylor Glacier, terminates in an ice-free valley. In the absence of East Antarctic outlet glaciers, late Quaternary ice sheets grounded in the Ross Sea infilled McMurdo Sound and spilled onto peripheral ice-free areas in the Transantarctic Mountains (Fig. 5). Westward-flowing ice lobes from these grounded ice sheets plugged the eastern ends of valleys that open onto the west coast of McMurdo Sound, including Taylor, Marshall, and Miers Valleys. Taylor Valley is particularly important, because it alone among these ice-free valleys contains an East Antarctic outlet glacier. Taylor Glacier, which terminates in the western end of the valley, originates in the peripheral McMurdo Dome on the East Antarctic plateau (Fig. 8). In turn, the small McMurdo Dome, which is controlled by subglacial bedrock, lies at the eastern end of an ice divide and flowline that extend inland to Dome Circe. Hence, the Taylor Valley geological record reveals the relative phasing of the McMurdo Dome and grounded ice sheets in the Ross Sea.

We now discuss three important drift sheets in the McMurdo Sound/Taylor Valley area. Two were deposited by ice sheets grounded in McMurdo Sound and one was deposited by an expanded Taylor Glacier. Ross Sea drift, the most prominent, represents the youngest infilling of McMurdo Sound by grounded ice and is described in detail by Stuiver *et al.* (1981, pp. 322–355). Figure 5 shows a reconstruction of the grounded Ross Sea ice sheet in McMurdo Sound. The ice-sheet limit in Figure 5

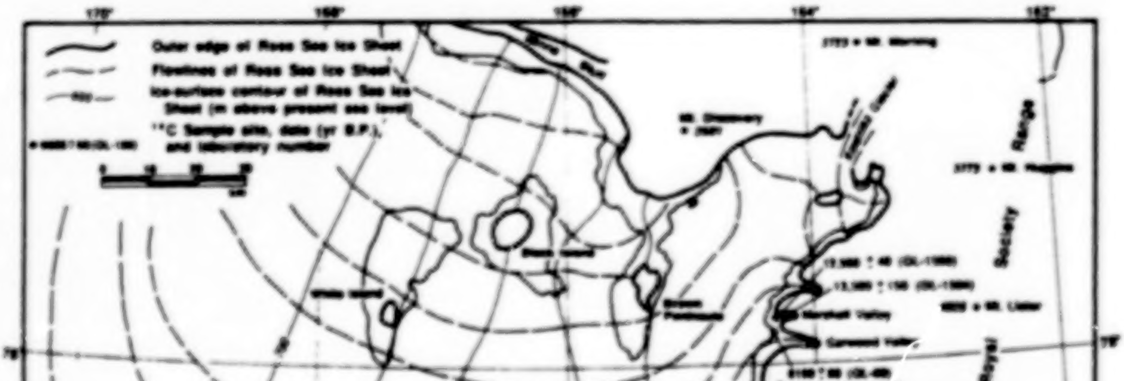

Outer edge of Ross Sea Ice Sheet
Flowlines of Ross Sea Ice Sheet
Ice-surface contour of Ross Sea Ice Sheet (m above present sea level)
¹⁴C Sample site, date (yr B.P.), and laboratory number
Society
Range
Royal
Marshall Valley
Garwood Valley
McMurdo
Sound
Ross Island
N

FIG. 5. Surface contours and flowlines of the grounded ice sheet that deposited Ross Sea drift in McMurdo Sound during late Wisconsin (stage 2) time. Adapted from Stuiver *et al.* (1981, Figs. 7–16).

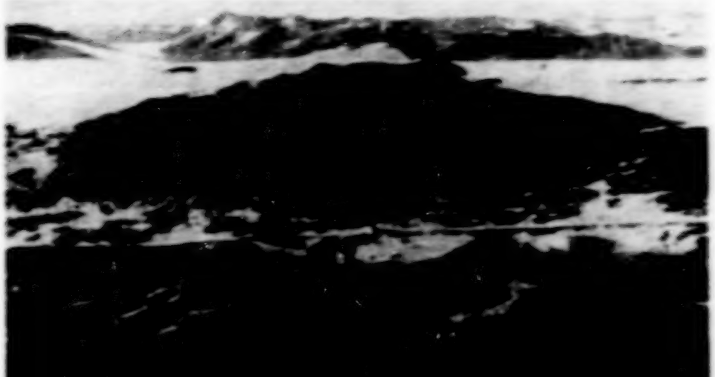


FIG. 6. Oblique aerial view of Ross Sea drift at the mouth of Taylor Valley. The lower part of the photograph shows sea ice on Explorers Cove.

comes from the geologic map depicted in Figure 7-3 of Stuiver *et al.* (1981). The surface contours are based on measured elevations of the drift limit, typical examples of which are shown in Figures 6 and 7. The former ice flowlines are perpendicular to the ice-surface contours and are consistent with the trend of debris bands and the distribution of distinctive erratics. Marshall drift, the second of these three drifts, is best exposed in Marshall Valley (Fig. 5) (Judd, 1986; Dagel *et al.*, 1989). The outer limit of Marshall drift is distal to and parallel with the Ross Sea drift limit along the coastal foothills and valleys of the Royal Society Range on the west coast of McMurdo Sound. From this configuration, as well as from the distribution of distinctive erratics, Dagel *et al.* (1989) inferred that Marshall drift represents the penultimate infilling of McMurdo Sound by a grounded ice sheet. The third important drift sheet in our discussion is Bonney drift, which occurs on the floor of Taylor Valley and in ice-free areas alongside Taylor Glacier. Bonney drift represents the penultimate advance of Taylor Glacier. It has correlatives alongside most alpine glaciers in Taylor Valley and elsewhere in the McMurdo Sound/Dry Valleys area. Figure 8 shows longitudinal profiles of the former Bonney and Ross Sea ice lobes in Taylor Valley based on the outer limits of the respective drift sheets.

From the reconstruction in Figure 5, we

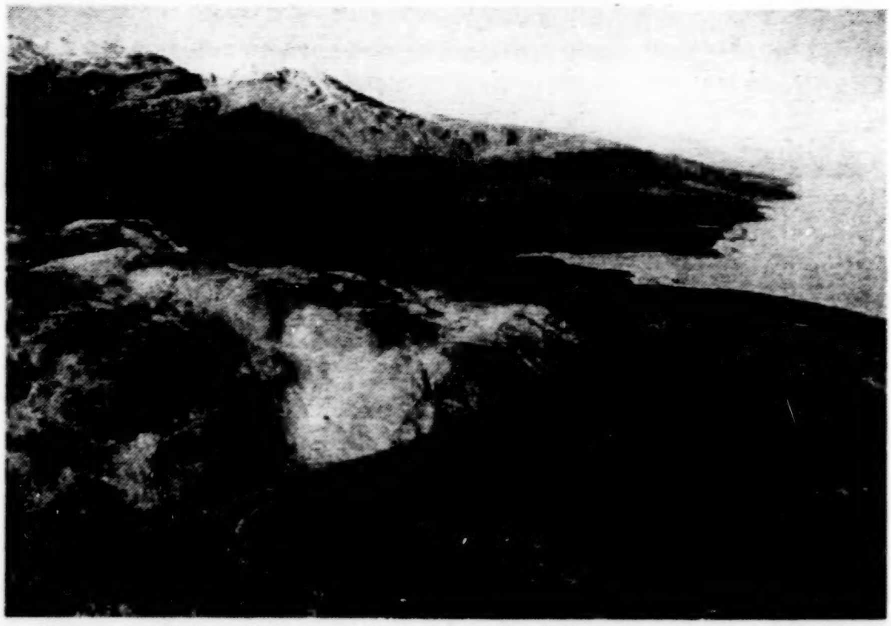


FIG. 7. Oblique aerial view of the upper limit of Ross Sea drift on the west coast of McMurdo Sound near the mouth of Miers Valley.

infer that the Ross Sea and Marshall drifts represent regional ice-sheet grounding, with ice flow from the eastern Ross Sea infilling McMurdo Sound in the absence of through-flowing East Antarctic outlet glaciers. In contrast, we attribute the Bonney advance to local climatic change rather than to ice-sheet grounding in the Ross Sea and McMurdo Sound. We do so for several reasons. First, the Bonney profile indicates that Bonney ice was not contiguous or contemporaneous with a grounded Ross Sea ice sheet in McMurdo Sound (Fig. 8). Hence, Bonney ice does not exhibit the enormous thickening near the Ross Sea shown by the outlet glaciers that drained directly into grounded ice in the Ross Embayment. Second, the Bonney advance was part of a regional expansion that simultaneously involved numerous local alpine glaciers as well as Taylor Glacier. This is distinctly different from the situation that occurred during the last grounding episode when local glaciers receded in the McMurdo Sound/Taylor Valley area (Stuiver *et al.*, 1981). Not only did Bonney drift antedate Ross Sea drift in central Taylor Valley, as shown by overlapping drifts (and hence longitudinal ice profiles) in central Taylor Valley (Fig. 8), but it directly abuts Ross Sea drift without the intervention of another drift sheet. This geometry almost certainly means that Bonney drift is intermediate in age between Ross Sea and Marshall drifts, which are contiguous on headlands along the west coast of McMurdo Sound. Such an age relationship implies that the Bonney advance was out-of-phase with the last two grounding episodes in McMurdo Sound and the adjacent Ross Sea. Numerical chronologies discussed below confirm these relative ages.

Figure 7-22 in Stuiver *et al.* (1981) shows two late Quaternary drift sheets at Terra Nova Bay in the northwestern Ross Embayment. Both drifts reflect thickening of at least the lower portions of local outlet glaciers, accompanied by ice-sheet grounding in Terra Nova Bay. Along the outer coast of Terra Nova Bay the upper surface of this grounded ice sheet reached 360 m during the younger grounding episode and 670–680 m during the older episode.

Chronology

Numerical chronology. To reconstruct the past behavior of the Ross ice drainage system, we correlate former ice profiles along the length of the Transantarctic Mountains by using our drift stratigraphy.

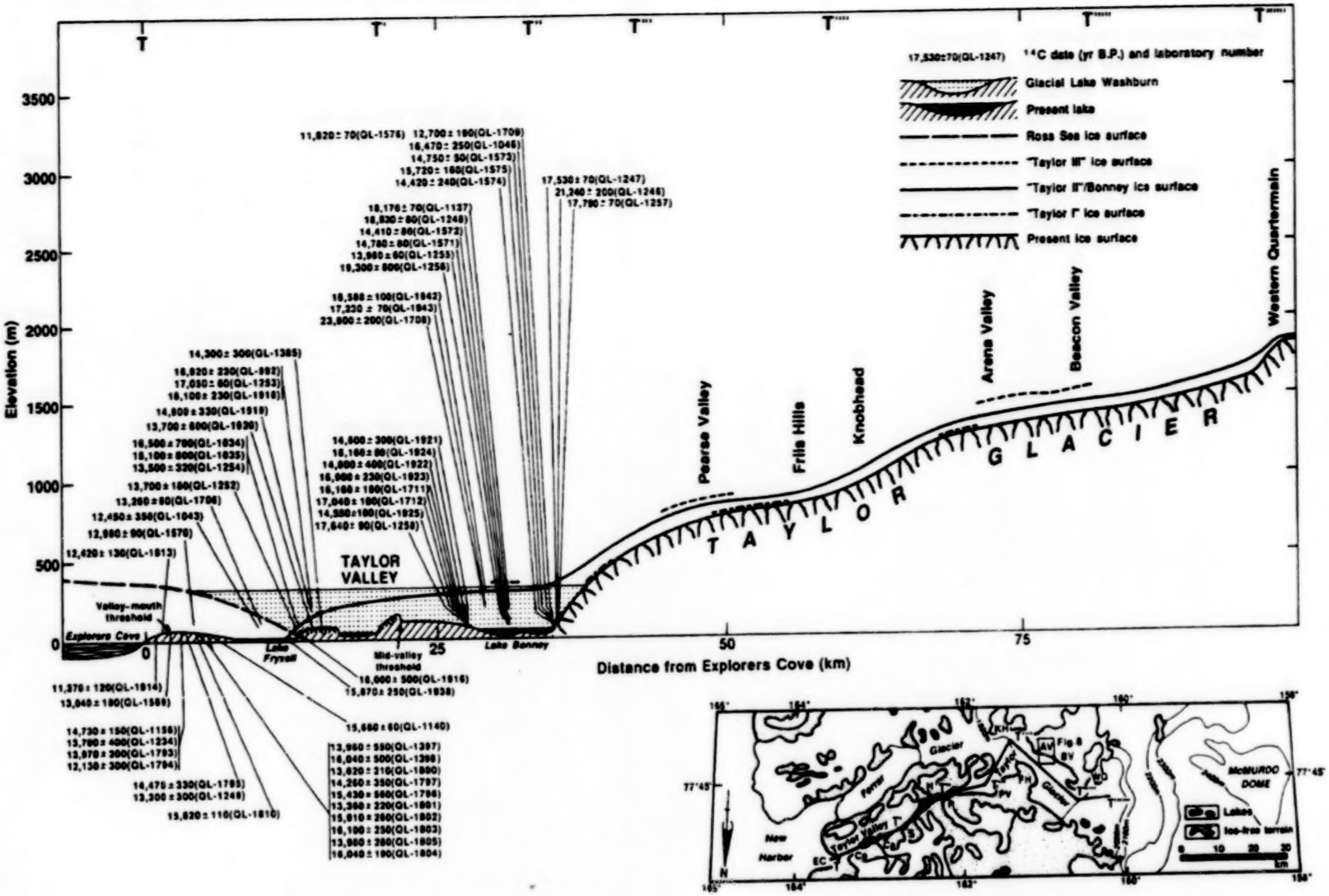
First, we develop a numerical chronology of Ross Sea, Marshall, and Bonney drifts in the McMurdo Sound/Taylor Valley area and give minimum ages for Britannia drifts in the Hatherton/Darwin Glacier area. We then apply this numerical chronology to a relative chronology based on drift sequence, surface morphology, and soil development. This relative chronology is the basis for correlation of drift sheets (and hence former longitudinal profiles) among outlet glaciers along the length of the Transantarctic Mountains. These former ice surface profiles allow reconstruction of the Ross ice drainage system at the height of late Wisconsin (Stage 2) glaciation.

An important part of our ^{14}C chronology comes from a unique lacustrine facies of Ross Sea drift. On the west coast of McMurdo Sound, the Ross Sea drift limit is sharp and commonly marked by a moraine ridge on headlands between ice-free valleys. In contrast, Ross Sea drift in valley floors has an irregular outer limit marked in places by near-horizontal terraces. This situation occurs because Ross Sea ice lobes in these valleys terminated in proglacial lakes. At the height of Ross Sea glaciation, interlocked glacial-and-lake-ice systems formed conveyer belts that transported drift from glacial lobes westward into the ice-free valleys on rafts of lake ice (Clayton-Greene, 1986; Clayton-Greene *et al.*, 1988a). During transit, fine-grained sediment passed through lake ice to form stratified lacustrine deposits on lake floors. Coarse sediment and clasts either dropped through fissures in lake ice or were deposited in moats at lake edges. The result was near-horizontal lake-edge terraces and a surficial mantle of valley-wall and lake-floor sand and gravel, some in the form of mounds and ridges. By this mechanism, most Ross Sea drift on valley floors is glacial lacustrine sediment deposited while ice tongues plugged valley mouths and supplied drift to contiguous lake ice. ^{14}C samples are plentiful in such glacial lacustrine drift in the form of blue-green algae and of carbonate beds.

A suite of ^{14}C dates comes from glacial lacustrine drift in Miers Valley, where a landward flowing Ross Sea ice lobe dammed Glacial Lake Trowbridge (Clayton-Greene, 1986). The ^{14}C dates pinpoint the age of the Ross Sea ice dam in several ways. First, Glacial Lake Trowbridge could not have existed in Miers Valley without an ice dam consisting of a Ross Sea glacier lobe in the mouth of Miers Valley. Further, most of the ^{14}C-dated carbonate beds are covered by erratics and gravel hummocks rich in McMurdo volcanics transported inland from the Ross Sea glacier lobe by the lake-ice raft (Clayton-Green, 1986). Hence, these deposits record the existence of a Ross Sea glacier lobe in the valley mouth. Fourteen dates of carbonate and one of blue-green algae all fall between 10,300 ± 900 yr B.P. (WK-718) and 22,950 ± 360 yr B.P. (WK-609), hence encompassing most of late Wisconsin time (Clayton-Greene *et al.*, 1988b).

Glacial Lake Washburn occupied the Bonney and Fryxell basins of Taylor Valley during the entire span of late Wisconsin time (Fig. 8) (Stuiver *et al.*, 1981). Two suites of ^{14}C dates of Glacial Lake Washburn lacustrine sediments show when a Ross Sea glacier lobe occupied eastern Taylor Valley. The first suite comes from blue-green algae in deltas of Glacial Lake Washburn. In the Fryxell basin, former lev-

FIG. 8. Longitudinal profiles of present and former glaciers in Taylor Valley. Position of profile is shown as T–T'''''' in Figure 12a. "Taylor I," "Taylor II," and "Taylor III" are temporary names for drift sheets alongside upper Taylor Glacier. "Taylor II" drift corresponds with Bonney drift in Taylor Valley. Inset map of McMurdo Dome is from Drewry (1980, 1982). The locations of the ^{14}C dates are schematic because they are projected onto the longitudinal profile from their position on the valley floor and walls. ^{14}C dates listed above the diagram designate deltas in Glacial Lake Washburn; those below the diagram designate Ross Sea glacial lacustrine drift.


T
T'
T''
T'''
T''''
T'''''
T''''''
17,530±70(QL-1247)
¹⁴C date (yr B.P.) and laboratory number
Glacial Lake Washburn
Present lake
Ross Sea ice surface
"Taylor III" ice surface
"Taylor II"/Bonney ice surface
"Taylor I" ice surface
Present ice surface
3500
3000
2500
2000
1500
1000
500
0
Elevation (m)
11,820 ± 70(QL-1576)
12,700 ± 190(QL-1709)
16,470 ± 250(QL-1046)
14,750 ± 50(QL-1573)
15,720 ± 160(QL-1575)
14,420 ± 240(QL-1574)
17,530 ± 70(QL-1247)
21,240 ± 200(QL-1246)
17,790 ± 70(QL-1257)
18,170 ± 70(QL-1137)
18,830 ± 80(QL-1248)
14,410 ± 80(QL-1572)
14,780 ± 80(QL-1571)
13,980 ± 60(QL-1255)
19,300 ± 800(QL-1256)
18,588 ± 100(QL-1942)
17,230 ± 70(QL-1943)
23,800 ± 200(QL-1708)
14,300 ± 300(QL-1385)
16,920 ± 230(QL-892)
17,050 ± 60(QL-1253)
18,100 ± 230(QL-1918)
14,800 ± 330(QL-1919)
13,700 ± 600(QL-1920)
16,500 ± 700(QL-1034)
15,100 ± 800(QL-1035)
13,500 ± 320(QL-1254)
13,700 ± 180(QL-1252)
13,280 ± 80(QL-1706)
12,450 ± 350(QL-1043)
12,880 ± 90(QL-1570)
12,420 ± 130(QL-1813)
14,800 ± 300(QL-1921)
18,180 ± 80(QL-1924)
14,800 ± 400(QL-1922)
16,900 ± 230(QL-1923)
16,180 ± 190(QL-1711)
17,040 ± 180(QL-1712)
14,550 ± 100(QL-1925)
17,840 ± 90(QL-1258)
TAYLOR VALLEY
Valley-mouth threshold
Explorers Cove
Lake Fryxell
Mid-valley threshold
Lake Bonney
Pearse Valley
Friis Hills
Knobhead
Arena Valley
Beacon Valley
Western Quartermain
TAYLOR GLACIER
0
25
50
75
Distance from Explorers Cove (km)
11,370 ± 120(QL-1914)
13,040 ± 190(QL-1569)
14,730 ± 150(QL-1156)
13,700 ± 400(QL-1234)
13,870 ± 300(QL-1793)
12,130 ± 300(QL-1794)
14,470 ± 330(QL-1795)
13,300 ± 300(QL-1249)
15,820 ± 110(QL-1810)
16,000 ± 500(QL-1916)
15,870 ± 250(QL-1938)
15,680 ± 60(QL-1140)
13,960 ± 550(QL-1397)
16,040 ± 500(QL-1398)
13,820 ± 210(QL-1800)
14,260 ± 350(QL-1797)
15,430 ± 560(QL-1796)
13,360 ± 220(QL-1801)
15,910 ± 260(QL-1802)
16,100 ± 250(QL-1803)
13,980 ± 280(QL-1805)
16,040 ± 190(QL-1804)
165°
164°
162°
160°
158°
77°45'
Ferrar
Glacier
Taylor
Glacier
New Harbor
Taylor Valley
EC
Cs
KH
AV
Fig 8
BV
FH
PV
MCMURDO DOME
Lakes
Ice-free terrain
0 10 20 30
km
N

els of Glacial Lake Washburn higher than the valley-mouth threshold, and in the Bonney basin higher than the mid-valley threshold, require a Ross Sea ice dam in eastern Taylor Valley. Details appear in Stuiver *et al.* (1981, pp. 345–355). Table 1 and Figures 8 and 9 display 32 ^{14}C dates of perched deltas in Taylor Valley that demonstrate such high lake levels. These ^{14}C dates all fall between 11,820 ± 70 yr B.P. (QL-1576) and 23,800 ± 200 yr B.P. (QL-1708). The second suite of ^{14}C dates comes from blue-green algae in glacial lacustrine drift on the eastern valley floor in the Fryxell and Explorers Cove basins. This drift was deposited by the conveyor-belt mechanism in which detritus from the Ross Sea ice lobe was transported into the Fryxell basin on the ice raft of Glacial Lake Washburn. Algae samples from such drift yielded 22 ^{14}C dates between 11,370 ± 120 yr B.P. (QL-1914) and 16,100 ± 250 yr B.P. (QL-1803) (Table 1 and Figs. 8 and 9). Together, the two suites of ^{14}C dates are consistent in indicating that a Ross Sea ice lobe plugged the mouth of Taylor Valley for at least most of Late Wisconsin (Stage 2) time between 23,800 and 12,450 yr B.P. In addition, the limited number of available ^{14}C dates of glacial lacustrine sediments suggests that the Ross Sea ice lobe projected far into the valley until about 16,000 yr B.P., that the Fryxell basin was largely ice free between 16,000 and 13,300 yr B.P. while the Ross Sea ice lobe terminated on the valley-mouth threshold, and that ice recession from the valley-mouth threshold (and hence decline of the ice-sheet surface in McMurdo Sound) was underway by 13,040 yr B.P.

Marshall drift is distal and parallel to Ross Sea drift along the western coast of McMurdo Sound. Its numerical age has been determined in Marshall Valley by 36 U/Th dates of carbonate lacustrine beds within volcanic-rich Marshall drift (Fig. 5) (Dagel *et al.*, 1988; Judd, 1986). All but four of these dates fall between 130,000 and 190,000 yr B.P. Hence, Marshall drift most likely corresponds with Stage 6 in the marine oxygen-isotope record.

There are two sources of numerical dating control on Bonney drift in Taylor Valley. The first is ^{14}C dates of perched deltas of Glacial Lake Washburn in the Bonney basin (Fig. 8, Table 1). These perched deltas are all younger than Bonney drift and therefore their late Wisconsin ^{14}C dates are minimum for this drift. It follows that Bonney drift predates late Wisconsin time and therefore is older than Ross Sea drift in the valley mouth. This confirms the relative ages of these two drifts that was inferred from their crosscutting relationships in central Taylor Valley (Fig. 8). From the areal distribution of ^{14}C-dated deltas in the Bonney basin (Fig. 8) it is clear that Glacial Lake Washburn, not Taylor Glacier, occupied much of the Bonney basin throughout late Wisconsin time. The areal distribution and ^{14}C ages of deltas adjacent to its current snout suggest that Taylor Glacier now occupies its maximum position since late Wisconsin time (with the exception of a slight fluctuation that in a few places has left a Holocene ice-cored moraine several meters from the current ice margin). This suggests strongly that the McMurdo Dome on the East Antarctic polar plateau surface inland from Taylor Valley (Fig. 8) was slightly lower than now during late Wisconsin time.

Direct numerical dates of Bonney drift come from carbonates deposited in small lakes dammed in middle Taylor Valley by an expanded Taylor Glacier during its Bonney advance. An extensive program is under way to date these carbonates by the U/Th method. Previously published dates (Hendy *et al.*, 1979) that we now interpret as relating to the Bonney drift are 75,000 ± 2600, 80,000 ± 2800, 92,000 ± 2000, 74,000 ± 1600, 95,000 ± 4500, and 98,000 ± 1700 yr B.P. Hence, Bonney drift most likely corresponds with Stage 5 in the marine oxygen-isotope record.

Farther south in the Transantarctic Mountains at Hatherton Glacier, ^{14}C dates discussed in detail by Bockheim *et al.*

TABLE 1. SELECTED ^{14}C DATES FROM THE TAYLOR VALLEY/MCMURDO SOUND AREA AND THE WESTERN ROSS EMBAYMENT

Laboratory number	^{14}C date (yr B.P.)	Location	Description	Reference
		^{14}C dates of Ross Sea drift in eastern Taylor Valley		
QL-1914	11,370 ± 120	Figure 8	Blue-green algae in Ross Sea glacial laustrine drift	This paper
QL-1794	12,130 ± 300	DO.	DO.	DO.
QL-1569	13,040 ± 190	DO.	DO.	DO.
QL-1249	13,300 ± 300	DO.	DO.	Stuiver *et al.* (1981)
QL-1801	13,360 ± 220 (AMS)	DO.	DO.	Denton *et al.* (1985)
QL-1800	13,620 ± 210	DO.	DO.	DO.
QL-1234	13,700 ± 400	DO.	DO.	Stuiver *et al.* (1981)
QL-1397	13,960 ± 550	DO.	DO.	Denton *et al.* (1985)
QL-1793	13,970 ± 300	DO.	DO.	This paper
QL-1805	13,980 ± 280	DO.	DO.	Denton *et al.* (1985)
QL-1797	14,260 ± 350	DO.	DO.	DO.
QL-1795	14,470 ± 330	DO.	DO.	This paper
QL-1156	14,730 ± 150	DO.	DO.	Stuiver *et al.* (1981)
QL-1796	15,430 ± 560	DO.	DO.	Denton *et al.* (1985)
QL-1140	15,660 ± 60	DO.	DO.	Stuiver *et al.* (1981)
QL-1810	15,820 ± 110	DO.	DO.	This paper
QL-1938	15,870 ± 250	DO.	DO.	DO.
QL-1802	15,910 ± 260 (AMS)	DO.	DO.	Denton *et al.* (1985)
QL-1916	16,000 ± 500	DO.	DO.	This paper
QL-1804	16,040 ± 190 (AMS)	DO.	DO.	Denton *et al.* (1985)
QL-1398	16,040 ± 500	DO.	DO.	DO.
QL-1803	16,100 ± 250 (AMS)	DO.	DO.	DO.
		^{14}C dates of perched deltas of Glacial Lake Washburn in Taylor Valley		
QL-1043	12,450 ± 350	Figure 8	Blue-green algae in Glacial Lake Washburn delta situated in Fryxell basin above the elevation of the valley-mouth threshold	Stuiver *et al.* (1981)
QL-1570	12,980 ± 90	DO.	DO.	Denton *et al.* (1985)
QL-1706	13,260 ± 80	DO.	DO.	DO.
QL-1254	13,500 ± 320	DO.	DO.	Stuiver *et al.* (1981)
QL-1252	13,700 ± 180	DO.	DO.	DO.
QL-1920	13,700 ± 600	DO.	DO.	This paper
QL-1919	14,800 ± 330	DO.	DO.	DO.
QL-1035	15,100 ± 800	DO.	DO.	Stuiver *et al.* (1981)
QL-1034	16,500 ± 700	DO.	DO.	DO.
QL-992	16,920 ± 230	DO.	DO.	DO.
QL-1253	17,050 ± 60	DO.	DO.	DO.
QL-1918	18,100 ± 230	DO.	DO.	This paper
QL-1385	14,300 ± 300	DO.	Blue-green algae in Glacial Lake Washburn delta situated in the Fryxell basin adjacent to the terminus of Canada Glacier	DO.
QL-1913	12,420 ± 130	DO.	Blue-green algae in Glacial Lake Washburn delta on the valley-mouth threshold	DO.
QL-1576	11,820 ± 70 —	DO.	Blue-green algae in Glacial Lake Washburn delta in the Bonney basin situated above the elevation of both the mid- and valley-mouth thresholds	Denton *et al.* (1985)

TABLE 1—*Continued*

Laboratory number	^{14}C date (yr B.P.)	Location	Description	Reference
QL-1709	12,700 ± 190	DO.	DO.	DO.
QL-1921	14,600 ± 300	DO.	DO.	This paper
QL-1573	14,750 ± 50	DO.	DO.	Denton *et al.* (1988)
QL-1922	14,800 ± 400	DO.	DO.	This paper
QL-1924	16,160 ± 80	DO.	DO.	DO.
QL-1711	16,160 ± 190	DO.	DO.	DO.
QL-1577	16,610 ± 70	DO.	DO.	Denton *et al.* (1985)
QL-1046	16,470 ± 250	DO.	DO.	Stuiver *et al.* (1981)
QL-1923	16,900 ± 230	DO.	DO.	This paper
QL-1943	17,230 ± 70	DO.	DO.	DO.
QL-1257	17,790 ± 70	DO.	DO.	Stuiver *et al.* (1981)
QL-1137	18,170 ± 70	DO.	DO.	DO.
QL-1942	18,580 ± 100	DO.	DO.	This paper
QL-1248	18,830 ± 80	DO.	DO.	Stuiver *et al.* (1981)
QL-1246	21,240 ± 200	DO.	DO.	DO.
QL-1708	23,800 ± 200	DO.	DO.	Denton *et al.* (1985)
QL-1255	13,980 ± 60	DO.	Blue-green algae in Glacial Lake Washburn delta situated in the Bonney basin below the elevation of the mid-valley threshold	Stuiver *et al.* (1981)
QL-1572	14,410 ± 80	DO.	DO.	This paper
QL-1574	14,420 ± 240	DO.	DO.	DO.
QL-1925	14,550 ± 100	DO.	DO.	DO.
QL-1571	14,780 ± 80	DO.	DO.	DO.
QL-1575	15,720 ± 160	DO.	DO.	DO.
QL-1712	17,040 ± 100	DO.	DO.	DO.
QL-1247	17,530 ± 70	DO.	DO.	Stuiver *et al.* (1981)
QL-1258	17,640 ± 90	DO.	DO.	DO.
QL-1256	19,300 ± 800	DO.	DO.	DO.
		^{14}C Dates from the Western Ross Embayment		
QL-174	7020 ± 60	Terra Nova Bay. See Figure 13	*Adamussium colbecki* valves from recent moraine at edge of Nansen Ice Sheet	Stuiver *et al.* (1981)
QL-141	5340 ± 50	Franklin Island. See Figure 13	Lowermost penguin remains in sea cliff exposure in platform on west side of Franklin Island	DO.
QL-71	6010 ± 70	Marble Point. See Figures 5 and 13.	Algae in emerged marine delta.	DO.
QL-72	6430 ± 70	DO.	*Adamussium colbecki* valves from emerged marine delta.	DO.
QL-96	6350 ± 60	DO.	DO.	DO.
L-627	5650 ± 150	DO.	Elephant seal in emerged beach deposits.	Nichols (1968)
QL-137	6050 ± 70	Explorers Cove, Taylor Valley. See Figures 5 and 13.	*Adamussium colbecki* valves from emerged marine deposits.	Stuiver *et al.* (1981)
QL-157	6150 ± 80	DO.	DO.	DO.
QL-191	6670 ± 200	DO.	*Adamussium colbecki* valves from within marine delta.	DO.
QL-993	8340 ± 120	DO.	Blue-green algae from lacustrine delta.	DO.
QL-1393	8900 ± 60	DO.	DO.	Denton *et al.* (1985)
QL-995	9860 ± 160	Lower Ferrar Valley. See Figures 5 and 13.	DO.	Stuiver *et al.* (1981)

TABLE 1—*Continued*

Laboratory number	^{14}C date (yr B.P.)	Location	Description	Reference
QL-1036	10,020 ± 40	DO.	DO.	DO.
Y-2399	9490 ± 140	West coast of McMurdo Sound. See Figure 5.	Blue-green algal mat in kettle lake on ice-cored Ross Sea drift.	DO.
QL-1590	12,500 ± 40	Marshall Valley. See Figure 5.	Blue-green algae from surficial lacustrine sediments on Ross Sea drift.	Dagel *et al.* (1989)
QL-1584	13,300 ± 150	DO.	Blue-green algae in Ross Sea glacial lacustrine drift.	DO.
QL-80	6190 ± 80	West coast of McMurdo Sound. See Figures 5 and 13.	Blue-green algal mat in kettle on ice-cored Ross Sea drift.	Stuiver *et al.* (1981)
QL-1126	6510 ± 50	McMurdo Ice Shelf. See Figures 5 and 13.	Marine shells from surface of McMurdo Ice Shelf.	Stuiver *et al.* (1981)
QL-166	6600 ± 60	DO.	DO.	DO.

Note. All dates are conventional dates (Stuiver and Polach, 1977) normalized on $\delta^{13}C = -25‰$. Corrections for seawater ^{14}C deficiency have not been made for dates of marine shells and penguin remains. AMS denotes accelerator mass spectrometry.

(1989) are consistent with the proposition that recession from the Britannia drifts was largely a late Wisconsin/Holocene phenomenon and that recession from Hatherton drift is Holocene in age. From geological considerations, we argued that the Britannia I and II drifts reflect the same major glaciation, with Britannia I drift representing a late readvance superimposed on ice recession from Britannia II drift (Bockheim *et al.*, 1989). If this argument is correct, the implication from the ^{14}C dates is that the Britannia II drift limit represents the late Wisconsin stand of Hatherton Glacier.

Relative chronology. To develop a relative chronology of drift sheets alongside Transantarctic outlet glaciers, we used a combination of geologic features (surface morphology, moraine pattern, and cross-cutting relationships), surface boulder weathering, and soil development (depth of staining, cohesion, visible salts, ghosts, maximum color development, morphogenetic salt stage, and weathering stage). First, we differentiated drift sheets in ice-free areas alongside Beardmore (Denton *et al.*, 1989), Hatherton (Bockheim *et al.*, 1989), and upper Taylor Glaciers (Fig. 10) (Bockheim, 1982). Second, we used soil development to determine relative ages of drifts among these outlet glaciers. We feel justified in this approach because drifts alongside these outlet glaciers are all composed of the same parent material (dolerite and sandstone gravel), occur near the polar plateau in an ultraxerous to xerous cold desert climate, and have similar topography (moraine crest) and elevation (1500–1800 m).

Table 2 and Figure 11 compare five soil morphologic features on drifts alongside upper Taylor, Hatherton, and Beardmore Glaciers. Depth of staining appears to be the most useful relative-age discriminator, followed by depth of coherence and depth of visible salts (Fig. 11). Although surface boulder weathering features are useful in distinguishing among drift sheets in a particular area, these features are not helpful in correlating drifts among areas. On the basis of analysis of variance there are few significant differences in soil properties on the Britannia I, Britannia II, and Beardmore drifts. There also are few differences in these properties on the Danum and Meyer drifts. "Taylor II"/Bonney drift has

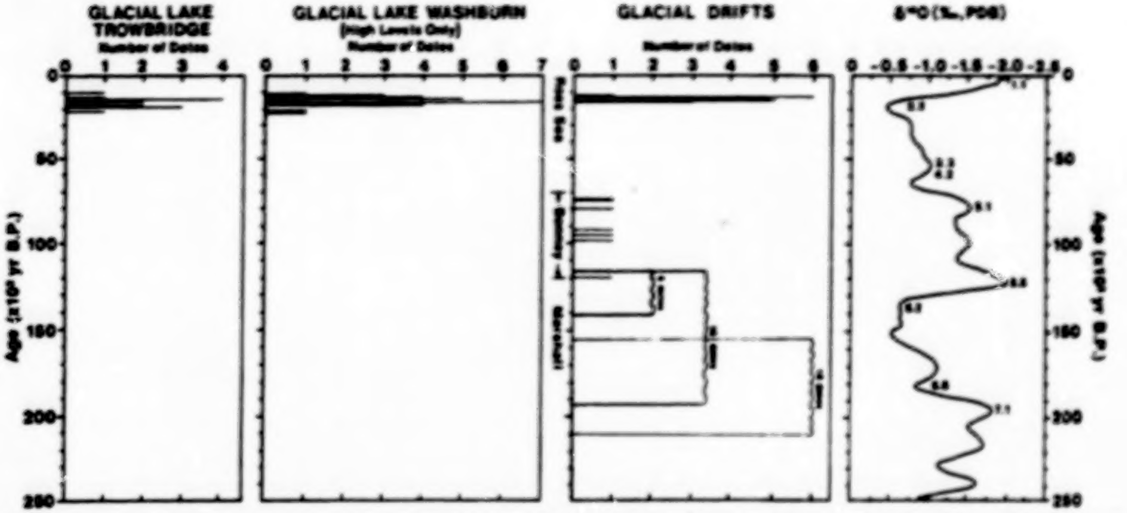


Fig. 9. Comparison of numerically dated glacial lacustrine sediments and glacial lake deltas in the McMurdo Sound/Taylor Valley area with marine oxygen-isotope stratigraphy (Prell *et al.*, 1986) that is taken as the standard for global climatic and ice-volume change. The ^{14}C dates for high levels of Glacial Lakes Trowbridge and Washburn and for Ross Sea drift are given in Table 1 and in Clayton-Greene *et al.* (1988b). The U/Th dates of Bonney drift are from Hendy *et al.* (1979). The U/Th dates for Marshall drift are given as three boxes that show the mean and standard deviation of numerous measurements on three carbonate beds within the drift. The values for these three beds are 129,000 ± 12,000, 155,000 ± 38,000, and 182,000 ± 28,000 yr B.P. (Judd, 1986; Dagel *et al.*, 1989).

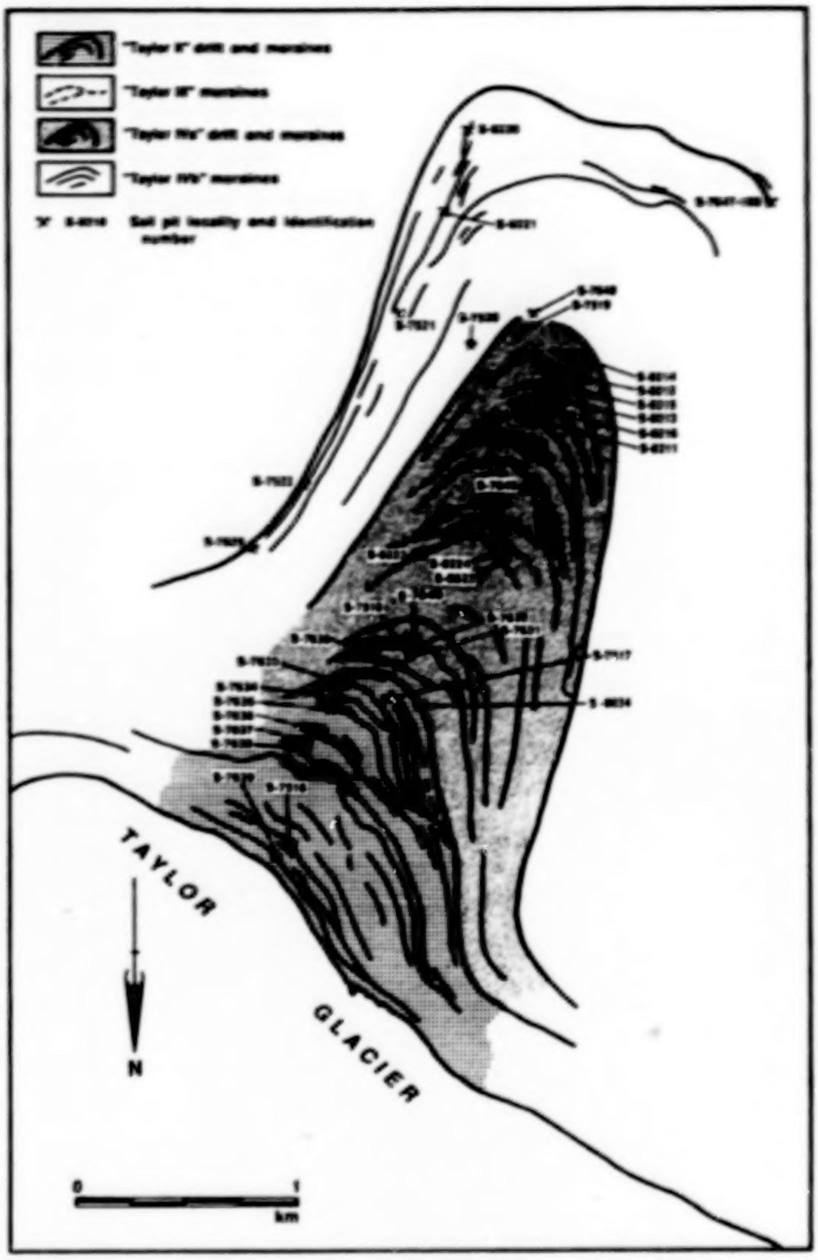

"Taylor II" drift and moraines
"Taylor III" moraines
"Taylor IVa" drift and moraines
"Taylor IVb" moraines
Soil pit locality and identification number
TAYLOR
GLACIER
N
0
1
km

FIG. 10. Drift sheets and moraines in Arena Valley beside upper Taylor Glacier. See Figure 8 for location. Map base from vertical aerial photograph. Drift names in quotes are temporary.

soil properties intermediate between the drifts of these two groups. Soil development is greatest on "Taylor III" drift.

These correlations are consistent with the geologic interpretations of the respective drift sheets. The Beardmore/Britannia and Meyer/Danum drift sets both produce longitudinal profiles of Transantarctic outlet glaciers that reflect grounded ice deep in the Ross Embayment. In contrast, the Bonney drift occurred between grounding episodes in McMurdo Sound and the adjacent Ross Embayment, and the "Taylor III" drift occurred prior to the grounding episode reflected by Meyer/Danum drift (Table 3).

Linkage of numerical and relative chronologies. Essential to our analysis of the Ross ice drainage system is the linkage of numerically dated drift sheets to the relative drift chronology derived from Figure 11. Bonney drift is prominent is both chronologies. Unfortunately, Ross Sea and

TABLE 2. MORPHOLOGICAL SOIL PROPERTIES ON DRIFTS IN THE BEARDMORE, HATHERTON, AND McMURDO SOUND/TAYLOR VALLEY AREAS OF THE TRANSANTARCTIC MOUNTAINS*

Drift	Isotopic stage	No. of profiles	Depth (cm)				C.D.E.*
			Staining	Coherence	Matrix salts	Ghosts	
Britannia I	2	9	2d	6b	0c	<1e	6d
Britannia II	2	10	5cd	5b	0c	7bcd	10c
Beardmore	2	41	6c	6b	0c	1e	7d
Bonney	5	11	8b	9b	0c	3cde	19a
Danum	6	10	10b	26a	8b	12b	13b
Meyer	6	30	10b	22ab	9ab	7c	10c
"Taylor III"	7	8	22a	29a	12a	21a	17a

* Color development equivalents (from Buntley and Westin, 1965).

* Values within a column followed by the same letter are not different at $P < 0.05$ based on Fisher's PLSD.

Marshall drifts cannot be tied into the relative chronology by using soil development because of distinct differences in drift composition among the pertinent drift sheets. Therefore, we use other arguments that involve geometric relations of drift sheets.

In Taylor Valley, Bonney drift is crosscut by Ross Sea drift (Fig. 8). This relative-age relationship is consistent with numerical ages for Bonney (Stage 5) and Ross Sea (Stage 2) drifts. Further, the lack of an intermediate-age drift implies that Bonney drift is younger than Marshall drift. Such a relative-age relation is supported by numerical ages for Ross Sea (Stage 2), Bonney (Stage 5), and Marshall (Stage 6) drifts. Soil development shows Britannia I and II drifts to be younger than Bonney drift (Fig. 11). This makes Ross Sea drift the most likely correlative of the Britannia drifts in the McMurdo Sound/Taylor Valley area. Such a correlation is consistent with the fact that Britannia and Ross Sea drifts are both taken to represent major grounding in this portion of the Ross Embayment. Further, from geologic considerations we argue that the Britannia drifts represent one major glacial event. Finally, available ^{14}C dates suggest that Britannia drift limits are late Wisconsin in age. For these reasons, we correlate Britannia II drift with the older portion of Ross Sea drift and Britannia I with the younger portion of Ross Sea drift. This means that Britannia I drift represents a readvance superimposed on general recession from the Britannia II drift limit. Such a correlation is discussed in detail elsewhere in this issue (Bockheim *et al.*, 1989).

The major weakness of correlating Britannia II and outer Ross Sea drift is that we

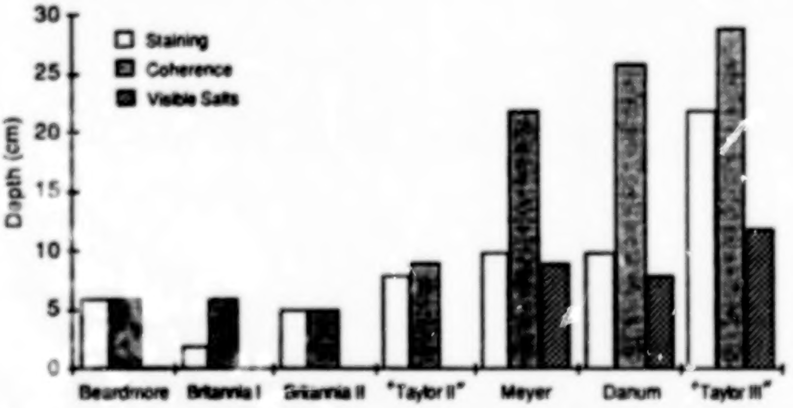


FIG. 11. Comparisons of soil morphologic properties among drift sheets alongside Taylor ("Taylor II"/"Taylor III"), Hatherton (Britannia I, Britannia II, and Danum), and Beardmore (Beardmore, Meyer) Glaciers (Fig. 1).

TABLE 3. SUGGESTED CORRELATIONS OF LATE QUATERNARY DRIFT SHEETS IN THE BEARDMORE GLACIER, HATHERTON GLACIER, AND MCMURDO SOUND/TAYLOR VALLEY AREAS OF THE TRANSANTARCTIC MOUNTAINS

Beardmore Glacier (Denton *et al.*, 1989)	Hatherton Glacier (Bockheim *et al.*, 1989)	McMurdo Sound/Taylor Valley	
		McMurdo Sound	Taylor Glacier
			"Taylor I" drift
Plunket drift	Hatherton drift		
	Britannia I drift		
Beardmore drift	Britannia II drift	Ross Sea drift	
			"Taylor II"/Bonney drift
Meyer drift	Danum drift	Marshall drift	
			"Taylor III" drift

have no control on soil development on sandstone–dolerite gravel drift numerically dated between isotope Stage 2 and Stage 6 in age. This weakness leaves open several alternative correlations. The first is that Britannia I drift correlates with outermost Ross Sea drift. This means that the Britannia II drift limit must represent a major grounding event that is younger than the Bonney advance and yet older and more extensive than Ross Sea grounding. A possible age for such a grounding event is that of isotope Stage 4. We think that this first alternative is implausible because it calls for a major grounding event near Hatherton Glacier that is not recognized in the McMurdo Sound/Taylor Valley area. Further, it is not in agreement with geologic considerations that Britannia I and II drifts cannot be separated in several areas alongside Hatherton Glacier (Bockheim *et al.*, 1989). The second alternative is that both Britannia drifts are older than Ross Sea drift and yet younger than Bonney drift. We think that this second alternative is unlikely because it calls for two major grounding episodes near Hatherton Glacier that go unrecognized in the McMurdo Sound record. In addition, it means that the extensive late Wisconsin grounding episode recorded in McMurdo Sound by the Ross Sea drift goes unrecorded near Hatherton Glacier.

We now expand our stratigraphy of late Quaternary drifts along the length of the Transantarctic Mountains. Table 3 illustrates our discussion. We start by comparing in Figure 9 our numerical drift chronology with the marine oxygen-isotope record taken as the standard of late Quaternary climate and ice-volume change on a global scale. In Figure 9 it is evident that the outer portion of Ross Sea drift is of late Wisconsin age and correlates with marine isotope Stage 2, the last global ice-sheet maximum. By our relative chronology, the outer limits of Beardmore and Britannia drifts are thought to correlate with outer Ross Sea drift from geological arguments given previously. On the basis of the agreement of field mapping results from the Beardmore Glacier area (Mercer, 1972; Denton *et al.*, 1988), we think it is very likely that Mercer's (1972) correlation of Reedy III and Beardmore III (our Beardmore) drifts is correct. We also follow Stuiver *et al.* (1981) in equating the "younger" drift at Terra Nova Bay with Ross Sea drift at McMurdo Sound. This correlation is based on surface weathering characteristics, morphologic drift form, limiting ^{14}C ages, and relative-age relationship to emerged Holocene beaches. In Terra Nova Bay and McMurdo Sound, these drifts both represent the youngest grounding episode, have well-preserved surface morphology, commonly exhibit sharp outer limits, have minimum ^{14}C ages of 7020–6600 yr B.P. (see below) for final ice recession, and are cut by emerged Holocene beaches (see below).

In Figure 9 Marshall drift corresponds with marine isotope Stage 6, representing the penultimate global glaciation. By our relative chronology, Marshall drift correlates with Danum drift at Hatherton glacier and Meyer drift at Beardmore Glacier. Finally, Figure 9 shows that Bonney drift correlates with marine isotope Stage 5.

Ice-Sheet Configuration

We use the upper limits of Ross Sea, Britannia II, Beardmore, Reedy III, and "younger" Terra Nova Bay drifts to reconstruct the late Wisconsin (Stage 2) configuration of the Ross ice drainage system. We assume our stratigraphy is correct. Another assumption is that these drift limits are strictly synchronous. This second assumption can only be tested by new numerical dating techniques. However, we recognize that the complex interactions of sea level and climate can lead to asynchronous drift limits during an individual glaciation.

We give two extreme reconstructions of the areal extent and surface elevations of the Ross ice drainage system at the height of late Wisconsin glaciation (Figs. 12a and 12b). Neither is based on ice-flow models, nor do we use seafloor sediments because they lack numerical chronology. Rather, both reconstructions are fixed by our outlet

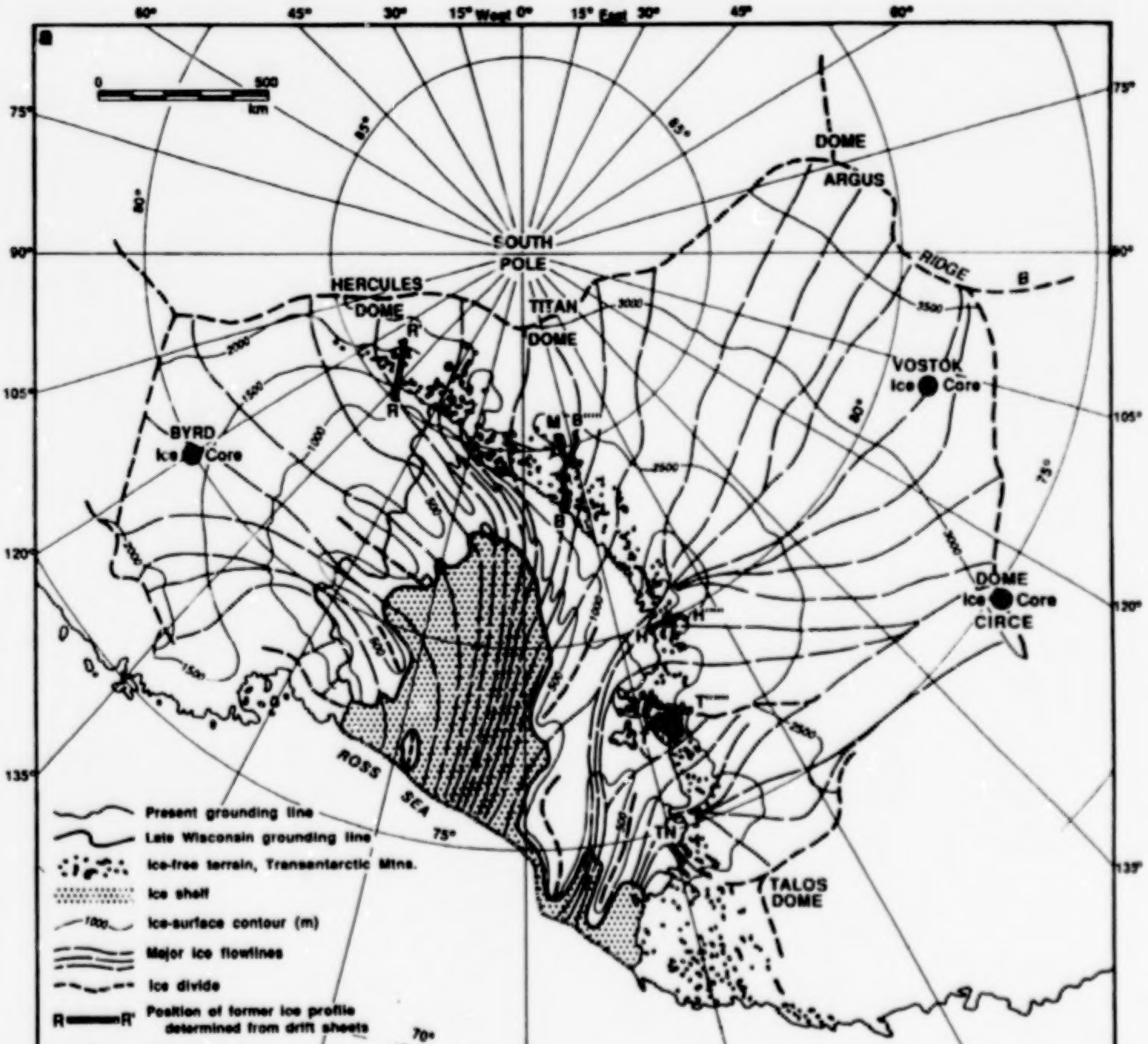


FIG. 12. Minimum (a) and maximum (b) reconstructions of grounded late Wisconsin ice in the Ross Embayment permitted by glacial geologic data from the Transantarctic Mountains. Isostatic compensation is not considered in these reconstructions. The position of the ice-shelf front shown in both diagrams is arbitrary. It is possible that shelf ice extended well offshore at the height of late Wisconsin glaciation. Lettered sections show positions of outlet glacier profiles in Figures 2, 3, 4, and 6.

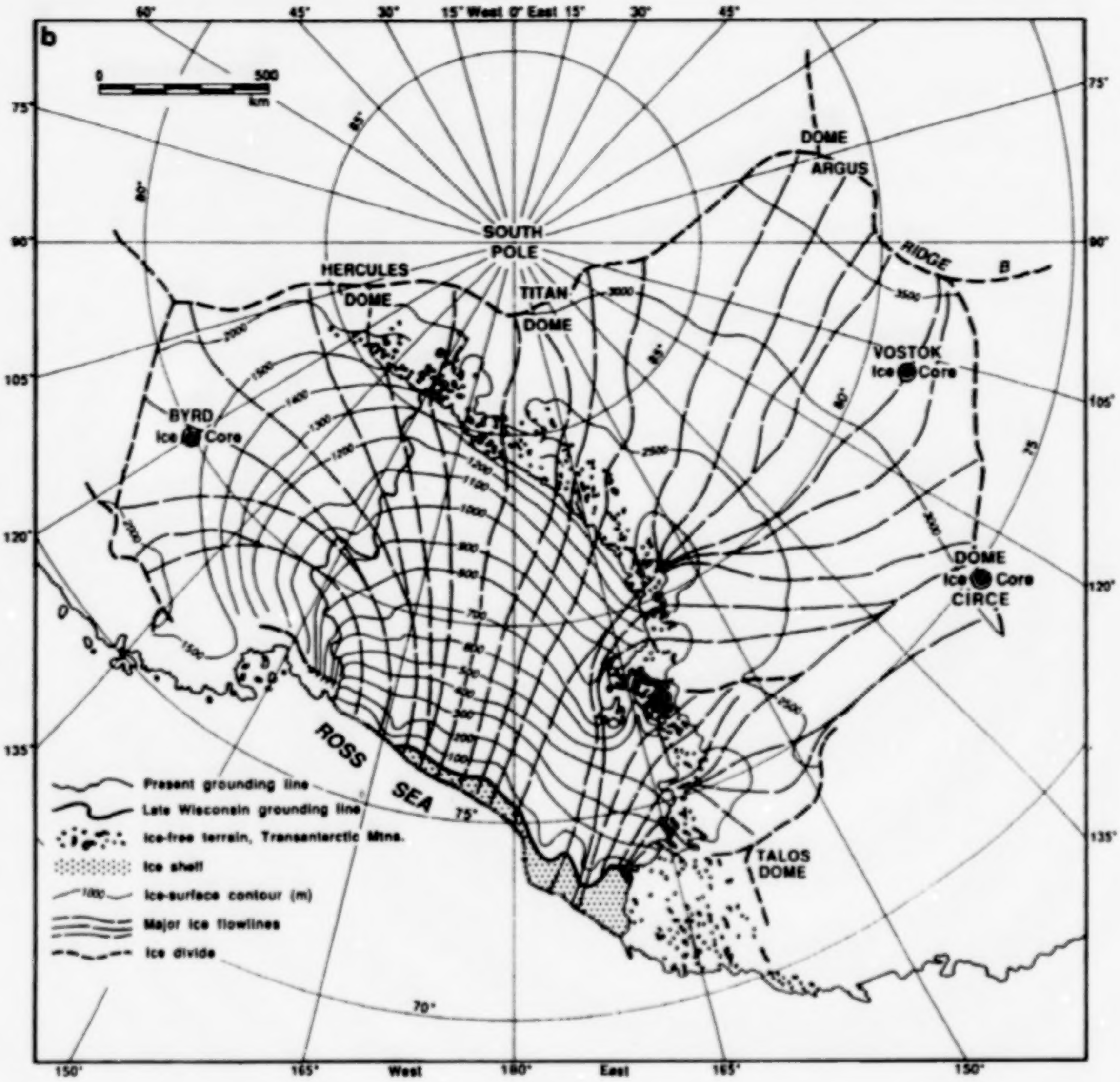


FIG. 12—*Continued*

glacier profiles, by the surface elevation of grounded ice in McMurdo Sound, by controlling ^{14}C dates on the extent of Taylor Glacier and, by implication, the height of the McMurdo Dome. As a result, both reconstructions depict grounded ice along the entire inner Ross Embayment and both show little change of the inland plateau surface adjacent to the Transantarctic Mountains between Taylor and Reedy Glaciers. The two reconstructions are identical in interior portions of the drainage system in East and West Antarctica. There we reproduce elevations similar to today's because of uncertainties about whether the inland surface increased or decreased slightly due to the interplay of sea-level and precipitation effects (this paper; Denton *et al.*, 1989; Alley and Whillans, 1984). The major difference lies in the areal extent of grounded ice in the outer Ross Embayment and is due to differing glaciological assumptions. The minimum reconstruction is based on the assumption that West-Antarctic-type ice streams persisted in the Ross Embayment at the last glacial maximum and hence that the surface slope of ice flowing from West Antarctica did not change significantly. Under this assumption, the Transantarctic Mountains data, particularly from Reedy and Beardmore Glaciers, limit seaward extent of the late Wisconsin grounding line. The resulting reconstruction shows a deep extension of shelf ice into the central Ross

Embayment (Fig. 12a). The maximum reconstruction is possible only if West-Antarctic-type ice streams did not exist in the Ross Embayment during late Wisconsin glaciation. This could occur if initial grounding-line advance was accompanied by coeval northward expansion of the Ross Ice Shelf to near the continental shelf edge, where it could ground extensively on submarine banks. The interior Ross Embayment could then fill by ice inflow from West Antarctica and from thickening Transantarctic outlet glaciers. The result, shown in Figure 12b, would be a thickening, slow-moving slab of grounded ice in the Ross Embayment that never achieved equilibrium conditions and that lacked through-flowing ice streams.

Climate

Late Wisconsin climate in the Ross ice drainage system is inferred from a combination of ice-core and geologic data. Jouzel *et al.* (1989) inferred from isotope records in the Dome Circe and Vostok ice cores (Fig. 1) that late Wisconsin full-glacial climate lasted from 25,000 to 15,000 yr B.P. Mean annual surface temperature is inferred to have been 9°C colder than now, and annual accumulation 50% less than now. Similar results come from the Byrd ice core in West Antarctica (Fig. 1) (Jouzel *et al.*, 1989).

Glacial geologic evidence from the Transantarctic Mountains bears on the interpretations of paleoclimate from isotope records of interior ice cores. One critical assumption of such interpretations is that the interior plateau ice surface was no higher than now at the late Wisconsin glacial maximum. Otherwise, parts or all of the isotope records may simply reflect changes in ice-surface elevation. Direct control on ice-sheet elevation could come from total gas content of the ice cores themselves (Lorius *et al.*, 1985), but the validity of this technique has been questioned (Paterson and Hammer, 1987). Our outlet-glacier profiles form an alternative control on interior ice elevations during late Wisconsin time, for they occur on strategic ice flowlines that lead back to interior domes and ice-core sites (Fig. 1). The Reedy III, Beardmore, and Britannia II profiles are consistent in indicating that the interior polar plateau near these flowlines probably did not rise more than 35–100 m during late Wisconsin (Stage 2) time. In fact, the glacial geologic evidence does not preclude a slight lowering along these flowlines, particularly inland of Beardmore Glacier. Taylor Glacier was no larger than now during late Wisconsin time. In fact, almost certainly its ice surface declined and its snout receded. This suggests little change, and probably a slight lowering, of the McMurdo Dome inland from Taylor Valley on the East Antarctic plateau. In summary, these glacial geologic data indicate little change of inland ice surfaces along critical ice flow lines that lead to major interior domes and pass near the important Vostok and Dome Circe ice cores, as well as little change of the McMurdo Dome. Hence, a significant rise in the interior ice surface can almost certainly be eliminated as a factor in interpreting the isotopic record at Dome Circe and Vostok in terms of temperature decrease during late Wisconsin time.

Is the reduced full-glacial precipitation inferred from ice cores consistent with the glacial geologic record in the Transantarctic Mountains? This record shows recession of Taylor Glacier and local alpine glaciers concurrent with full expansion of grounded Ross Sea ice during late Wisconsin time (Stuiver *et al.*, 1981). Further, Rutkowski Glacier in the Dominion Range was behind its present terminal position when Beardmore Glacier thickened to its late Wisconsin limit (Denton *et al.*, 1989). Finally, the plateau surface along flowlines inland of our outlet-glacier profiles showed little change, and may have declined, at the late Wisconsin maximum. Together, these glacial events can all be explained by reduced late Wisconsin precipitation in the Transantarctic Mountains, an interpretation consistent with the inferences drawn from isotope records in ice cores.

There is, however, one aspect of the glacial geologic record that may be at variance with the paleoclimatic interpretation from ice cores. From our previous discussion, it is obvious that large lakes formed in the Dry Valleys during late Wisconsin time. For example, Glacial Lake Washburn occupied Taylor Valley (Stuiver *et al.*, 1981; Denton *et al.*, 1988) (Fig. 8) and Glacial Lake Trowbridge existed in Miers Valley (Clayton-Greene *et al.*, 1988b). Our preliminary mapping results in other ice-free valleys indicate that such lakes were widespread and not restricted solely to Taylor and Miers Valleys.

We do not yet know why such lakes existed during the last glacial maximum. Several explanations are possible and each should be tested thoroughly. The first is that subglacial outflow from beneath Ross Sea ice lobes fed these lakes (Stuiver *et al.*, 1981). We no longer favor this alternative. A second explanation is that input came solely from surface glacier melt. The increased volume of meltwater associated with higher lake levels could have come from increased intensity of melt, or from a longer melt season, or from both factors. There are several variations on this scenario. The increased volumes of meltwater necessary for Glacial Lake Trowbridge and Glacial Lake Washburn in the Fryxell basin could have come solely from the surface of Ross Sea ice lobes projecting into Miers and Taylor Valleys. But such an explanation cannot apply to the Bonney basin of Glacial Lake Washburn. Here late Wisconsin lake levels rose to the mid-valley threshold without any overflow from the Fryxell basin. This suggests that increased volume of meltwater from local glaciers, as well as from Ross Sea ice lobes, is almost certainly an important contributing factor to high late Wisconsin lake levels. In such a case, the increased meltwater input could have resulted simply from warmer late Wisconsin temperatures (Denton *et al.*, 1985). Such postulated warming could have been widespread, in which case there is a serious discrepancy with temperature inferences drawn from all three major ice cores in the Ross ice drainage system. Such widespread warming also disagrees with our inferences that late Wisconsin retraction of local glaciers resulted from widespread aridification in the Transantarctic Mountains. Or the temperature increase could have been confined to the Dry Valleys. Clayton-Greene *et al.* (1988b) suggested that one local cause of such warming might have been a decrease in katabatic winds. We do not find this suggestion appealing, for we think that continued katabatic winds during late Wisconsin time were necessary to ablate Ross Sea glacier lobes flowing into ice-free valleys.

An alternative solution to the glacial-lake problem depends on the sensitivity to snow cover of meltwater production in the ice-free valleys. For example, meltwater production in the Dry Valleys can be decreased dramatically by late spring snowstorms (Chinn, 1981). Widespread grounded and shelf ice in the Ross Sea, along with lower ice-age temperatures, could have promoted the aridification in the Dry Valleys and elsewhere along the Transantarctic Mountains inferred from shrinkage of small glaciers and little change in the inland plateau surface. Coastal storms would no longer have penetrated the Dry Valleys. Such intense aridification would have resulted in little or no snow cover in the Dry Valleys through the entire sunlight season. Together with the low-albedo ablation zone of the Ross Sea ice sheet in McMurdo Sound and in eastern Taylor Valley (due to surface volcanic debris), this aridification could have decreased the regional albedo and, in the absence of snowfall, the length of the melt season. This explanation for increased lake levels is not in conflict with lower late Wisconsin regional temperatures and accumulation.

HOLOCENE STAGE

General Statement

Glacial geologic evidence indicates that grounded ice receded from the western Ross Embayment at the end of late Wisconsin and beginning of Holocene time. Simultaneous expansion occurred for local alpine glaciers and for Taylor Glacier. Isotopic records from the Vostok and Dome Circe ice cores are interpreted in terms of climatic warming and increased accumulation beginning abruptly about 15,000 yr B.P. and continuing into early Holocene time (Lorius *et al.*, 1984, 1985; Jouzel *et al.*, 1987, 1989). We review here our current knowledge of the glacial history of the Transantarctic Mountains during the late Wisconsin/early Holocene transition.

Radiocarbon Chronology

Ross Sea and equivalent drifts. In the previous section we argued that the distal portions of Ross Sea, Britannia II, Beardmore, Reedy III, and the "younger" drift at Terra Nova Bay date to the late Wisconsin maximum. We now review available ^{14}C dates of the proximal portions of these drift sheets. Such dates give the timing of grounded ice recession from the western and southwestern Ross Embayment. We start in the north at Terra Nova Bay.

In the Terra Nova Bay region *Adamussium colbecki* shells from a recent moraine at the edge of the floating Nansen Ice Sheet afforded an age of 7020 ± 60 yr B.P. (QL-174) (Table 1; Fig. 13). This date is minimum for grounded ice recession from the sample site and hence minimum for the "younger" drift.

On the west coast of McMurdo Sound, several ^{14}C dates pertain to proximal Ross

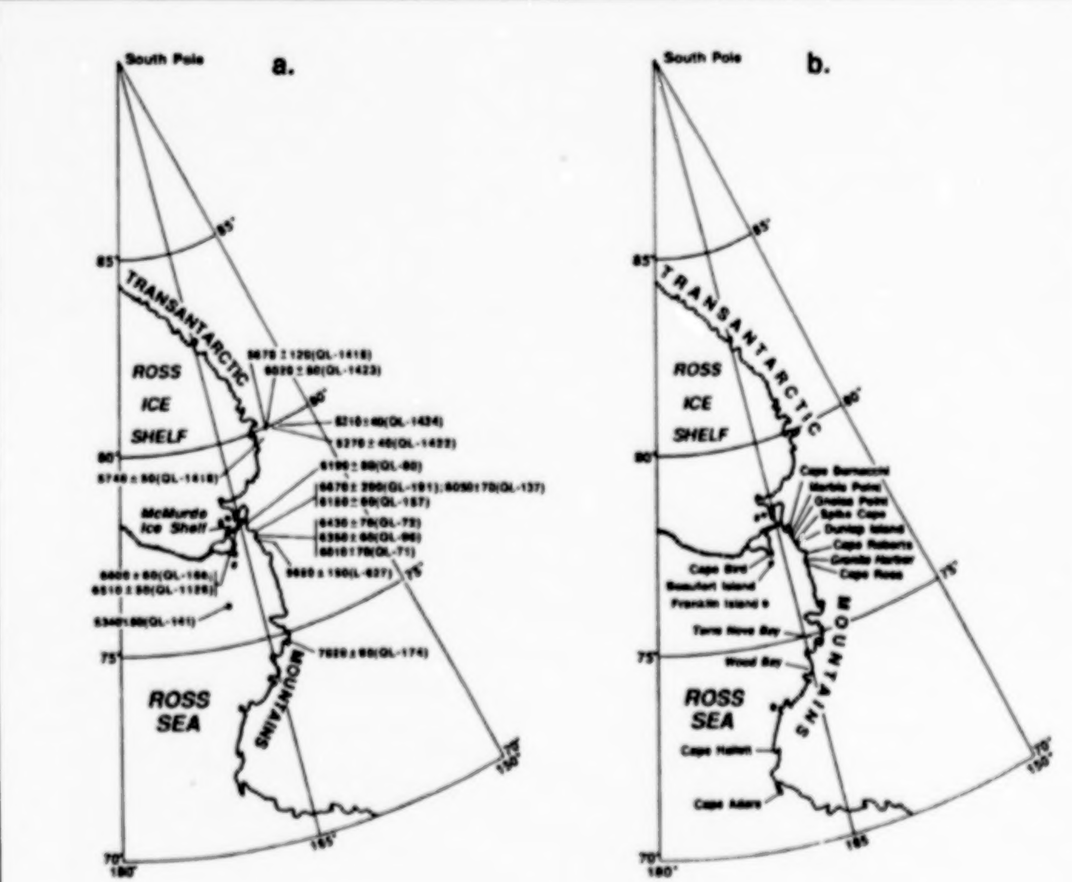


FIG. 13. Selected ^{14}C dates of late Wisconsin/Holocene recession of grounded ice from the western Ross Embayment (a); locations of raised beaches in the western Ross Embayment (b).

Sea drift (Table 1; Fig. 7). In Taylor Valley a date of 13,040 ± 190 yr B.P. (QL-1569) for blue-green algae in glacial lacustrine sediments affords a minimum age for recession of the Ross Sea glacial lobe from the valley-mouth threshold near Explorers Cove. A similar sample on the seaward slope of this valley-mouth threshold gives a minimum age for ice recession from the sample site of 11,370 ± 120 yr B.P. (QL-1914). A lacustrine delta near the top of this threshold dates to 12,420 ± 130 yr B.P. (QL-1913) (Fig. 8), giving a minimum age for ice recession from the sample site. Lacustrine deltas on the same seaward slope of the threshold record a lake dammed between the threshold and the retreating Ross Sea ice. ^{14}C dates of blue-green algae in these deltas indicate that grounded Ross Sea ice still occupied McMurdo Sound near Taylor Valley as late as 8900 ± 60 yr B.P. (QL-1393) and 8340 ± 120 yr B.P. (QL-993) (Fig. 5).

A complex of small deltas rests on Ross Sea drift on the south wall of Ferrar Glacier valley near McMurdo Sound. These deltas postdate the maximum of Ross Sea ice extent and hence afford minimum ^{14}C ages for Ross Sea drift (Table 1; Fig. 5). Dates of two deltas at 43 and 62 m elevation are 10,020 ± 40 yr B.P. (QL-1036) and 9860 ± 160 yr B.P. (QL-995), respectively. Along the west coast of McMurdo Sound near the Royal Society Range, minimum ages for lowering of the Ross Sea ice surface to less than 100 m above present sea level are 9490 ± 140 yr B.P. (Y-2399) and 12,500 ± 40 yr B.P. (QL-1590) on samples of blue-green algae. Finally, a date of 6190 ± 80 yr B.P. (QL-80) on blue-green algae is minimum for ice recession to the present coastline.

Further south in the Transantarctic Mountains, ^{14}C dates of blue-green algae deposited in small lakes perched beside the receding Hatherton Glacier afford minimum ages for ice-surface lowering (Figs. 4 and 13; Bockheim *et al.*, 1989). They indicate lowering of the Hatherton Glacier surface to near its present position by 5210 ± 40 yr B.P. (QL-1424) in its upper reaches, 5670 ± 120 yr B.P. (QL-1415) in its middle reaches, and 6020 ± 50 yr B.P. (QL-1423) in its lower reaches. Lower Darwin Glacier near the Ross Ice Shelf achieved its current surface elevation by 5740 ± 50 yr B.P. (QL-1418).

Emerged marine deposits. Emerged beach deposits occur along the west coast of the Ross Embayment from Cape Bernacchi in McMurdo Sound northward to Cape Adare, as well as at Cape Bird and on Beaufort and Franklin Islands in the Ross Sea (Fig. 13). Locations of emerged beaches shown in Figure 13 are from Nichols (1968), Stuiver *et al.* (1981), and Mabin (1986). All these emerged beaches are 35 m or lower in elevation (Nichols, 1968; Stuiver *et al.*, 1981; Mabin, 1986). These beaches are taken to represent isostatic rebound consequent on recession of grounded ice from the western Ross Embayment. Although these emerged beaches are consistent with the former presence of grounded ice in the western Ross Embayment, they are not a straightforward indicator of former ice thickness for two reasons. First, the beaches are largely undated and thus cannot afford isobases of uplift. Second, the highest beach in an individual area may simply reflect uplift since recession of floating shelf ice rather than of grounded ice (Denton *et al.*, 1988).

There are few ^{14}C dates associated with these beaches. One relates directly to a beach at Marble Point (5650 ± 150 yr B.P.: L-627) (Nichols, 1968). Others are of associated deltas at Marble Point (6010 ± 70 yr B.P., QL-71; and 6430 ± 70 yr B.P., QL-72) and bedded colluvium at Franklin Island (5340 ± 50 yr B.P., QL-141) (Table 1; Figs. 5 and 13).

Emerged marine deposits occur at the mouth of Taylor Valley at the edge of Explorers Cove despite the absence of well-defined beaches. Numerous ^{14}C dates of *A. colbecki* valves come from these deposits or from deep within a marine delta at the valley mouth. These dates are minimum for incursion of marine waters into Explorers Cove in early Holocene time. The oldest such dates are 6050 ± 70 yr B.P. (QL-137), 6150 ± 80 yr B.P. (QL-157), and 6670 ± 200 yr B.P. (QL-191). Together with the ^{14}C ages of lacustrine deltas on the seaward slope of the valley-mouth threshold, these dates indicate recession of grounded Ross Sea ice from this sector of McMurdo Sound between 8340 and 6670 yr B.P.

McMurdo Ice Shelf. The McMurdo Ice Shelf, an extension of the Ross Ice Shelf, now covers southern McMurdo Sound and contains evidence that bears on deglacial history of grounded ice from the Ross Embayment (Stuiver *et al.*, 1981). Some debris bands on the shelf surface are remnant from the Ross Sea Ice Sheet (Stuiver *et al.*, 1981); they show that the ice shelf formed in part by thinning of the grounded Ross Sea Ice Sheet in McMurdo Sound. Other debris bands formed in Holocene time by basal freezing and seaward flow. One such band that extends northward from Black Island to the shelf edge displays progressively greater ^{14}C ages of shells and barnacles from south to north. The oldest samples at the northern tip of this band afforded ages of 6510 ± 50 yr B.P. (QL-1126) and 6600 ± 60 yr B.P. (QL-166) (Table 1; Figs. 5 and 13) (Stuiver *et al.*, 1981). Hence, Holocene grounding-line recession of the Ross Sea Ice Sheet had reached Black Island by 6600 yr B.P. The McMurdo Ice Shelf extended farther north until recently, thereby explaining the absence of emerged beaches south of Cape Bird on Ross Island and south of Cape Bernacchi along the west coast of McMurdo Sound (Stuiver *et al.*, 1981).

Taylor Glacier and alpine glaciers. Taylor Glacier and alpine glaciers in Taylor Valley both occupy their maximum positions since late Wisconsin time, with the exception of a minor fluctuation shown by scattered ice-cored moraines of Holocene age. The evidence comes largely from ^{14}C-dated perched deltas of Glacial Lake Washburn (Fig. 8) near Taylor Glacier that were discussed previously. Rhone Glacier is now overriding deltas dated to 12,700 ± 190 yr B.P. (QL-1709) and 16,470 ± 250 yr B.P. (QL-1046); Hughes Glacier terminus is close to deltas dated to 18,170 ± 70 yr B.P. (QL-1137) and 18,830 ± 80 yr B.P. (QL-1248); and Canada Glacier terminus is adjacent to a delta dated to 14,300 ± 300 yr B.P. (QL-1385). Commonwealth Glacier rests on Ross Sea glacial lacustrine drift dated nearby to 15,660 ± 60 yr B.P. (QL-1140); 13,970 ± 300 yr B.P. (QL-1793); 12,130 ± 300 yr B.P. (QL-1794); 14,470 ± 330 yr B.P. (QL-1795); 14,730 ± 150 yr B.P. (QL-1156); and 13,700 ± 400 yr B.P. (QL-1234). That this situation is not unique to Taylor Valley is shown by several alpine glaciers along the west coast of McMurdo Sound that have advanced into Ross Sea drift or over emerged Holocene beaches.

Glacial lakes. In Taylor Valley high levels of Glacial Lake Washburn in the Bonney basin persisted above the mid-valley threshold until 12,700 ± 190 yr B.P. (QL-1709), and perhaps until 11,820 ± 70 yr B.P. (QL-1576). In the Fryxell basin they persisted above the valley-mouth threshold until at least 12,450 ± 350 yr B.P. (QL-1043) (Fig. 8). Hence, we conclude that a Ross Sea glacier lobe plugged eastern Taylor Valley until this time. We know from ^{14}C dates given earlier that the Ross Sea glacier lobe in eastern Taylor Valley had retreated from the crest of the valley-mouth threshold by 13,040 yr B.P. and from its seaward slope by 11,370 yr B.P. (Fig. 8). Nevertheless, higher-than-present lake levels persisted in the Fryxell basin as late as 9200 ± 40 yr B.P. (QL-1142) (Stuiver *et al.*, 1981), although such levels were much lower than the valley-mouth threshold. One probable water source for this late-lingering lake was westward overflow through a channel on the valley-mouth threshold from a lake trapped between the threshold and a grounded Ross Sea ice sheet in Explorers Cove. Such a trapped lake existed until at least 8340 ± 120 yr B.P. (QL-993), the age of the lowest lacustrine delta on the seaward slope of the threshold. This ice-dammed lake had ceased to exist by 6670 ± 200 yr B.P. (QL-191), the age of the oldest marine shells in Explorers Cove. Hence, we infer that grounded Ross Sea ice cleared Explorers Cove between 8340 and 6670 yr B.P. (Fig. 13).

The most striking feature highlighted by the ^{14}C dates in Taylor Valley is that high lakes in the Fryxell and Bonney basins were common during the entire span of late Wisconsin time but were almost absent during Holocene time. The exception is the late-lingering, but relatively low, lake that lasted in Fryxell basin until 9200 yr B.P.

Late Wisconsin/Early Holocene Ice-Sheet Changes

In summary, we infer from our data that the grounded Ross Sea ice sheet receded from the western Ross Embayment in late Wisconsin and early Holocene time. Radiocarbon dates in Marshall and Taylor Valleys indicate that Ross Sea glacier tongues had receded eastward nearly into McMurdo Sound by 12,500–13,040 yr B.P., but ^{14}C dates from Ferrar and Taylor Valleys imply that grounded ice lingered in western McMurdo Sound until 10,020 to 8340 yr B.P. Grounded Ross Sea ice had cleared the coast of the western and southern Ross Embayment by 7020 yr B.P. at Terra Nova Bay (Stuiver *et al.*, 1981); 6430, 6670, 6600, 6190, and 7750 yr B.P. at McMurdo Sound (Stuiver *et al.*, 1981); and 5210, 5670, 6020, and 5750 yr B.P. at Hatherton Glacier (Bockheim *et al.*, 1989).

Our geologic evidence shows strikingly different behavior of the plateau surface of the East Antarctic Ice Sheet inland of the Transantarctic Mountains. No massive decrease in surface elevation occurred here in late Wisconsin/Holocene time comparable with recession of grounded ice from the Ross Embayment. The Reedy III, Beardmore, and Britannia II profiles show that the decline of the upper reaches of the Reedy, Beardmore, and Hatherton Glaciers during the late Wisconsin/Holocene transition was only 30–100 m. In fact, these ice profiles even allow the possibility that the plateau surface along inland flowlines could be higher now than that during late Wisconsin time. In accord with this, Taylor Glacier and local alpine glaciers have expanded so that they are now at, or very close to, their maximum positions since late Wisconsin time. Likewise Rutkowski Glacier, which drains an independent ice cap on the Dominion Range near the upper reaches of Beardmore Glacier, has advanced at least 2.5 km over Beardmore drift and is now at its most robust configuration since late Wisconsin time.

Climate

Holocene surface climate in the Ross ice drainage system is inferred from a combination of ice-core and geologic data. Isotope records from the Vostok and Dome Circe ice cores are taken to show a marked warming that terminated late Wisconsin glacial climates. The ice-core chronology, based on an ice-flow model tied to accumulation, places the beginning of this climatic warming at about 15,000 yr B.P. The full amplitude of temperature change across the late Wisconsin/Holocene transition is taken to be 9°–10°C, and the Holocene accumulation rate is interpreted to be about twice the full-glacial value (Jouzel *et al.*, 1989).

Glacial geologic evidence bears on these interpretations of paleoclimates from interior ice cores. First, our outlet glacier profiles show little change, and permit even a slight rise, of the inland plateau. The strong implication is that the isotopic change in interior ice cores across this same transition reflects widespread climatic change and not simply massive lowering of the interior plateau ice surface. The expansion of Taylor Glacier and local alpine glaciers, as well as the advance of Rutkowski Glacier at the head of Beardmore Glacier, during the late Wisconsin/Holocene transition is best explained by increased accumulation in the Transantarctic Mountains and on the polar ice plateau just inland (Denton *et al.*, 1971, 1989). This, too, is in accord with inferences on increased interior accumulation drawn from ice cores.

A problem again arises in interpreting glacial lake history in the Dry Valleys in terms of a rise in temperature and accumulation through the late Wisconsin/Holocene transition. Extensive late Wisconsin lakes were succeeded by limited Holocene lakes generally even less extensive than those of today. Why would lakes decrease in both area and volume, if the climate warmed and precipitation increased? There are at least three explanations for this observed history of glacial lakes. First, the meltwater source from Ross Sea ice lobes disappeared with recession of grounded ice from McMurdo Sound. That this is partly responsible is shown by the lake-level history in eastern Taylor Valley. However, it does not explain the disappearance of lakes in the Bonney basin. The second explanation is that lake-level variations reflect strictly the effect of summer temperatures on glacier melt and hence on lake-level variation. This scenario disagrees with regional temperature changes inferred from ice cores. A third explanation involves both the recession of Ross Sea ice from McMurdo Sound and the decreasing aridity of the Dry Valleys through the late Wisconsin/Holocene transition. Ross Sea ice recession would remove a low-albedo, debris-laden ablation ice surface from McMurdo Sound, while simultaneously opening the western Ross Sea and removing a barrier to the penetration of storms. One result could be increased spring and summer albedo in McMurdo Sound from replacement of dirty glacial ice by clean sea ice and in the Dry Valleys from sporadic snowstorms. The overall effect could be to decrease the length of the melt season, while at the same time removing a meltwater source. These mechanisms could produce the observed lake-level history by climatic changes consistent with those inferred from ice cores.

DISCUSSION

The first recognition of a former grounded great ice sheet in the Ross Embayment came early in this century (Scott, 1905; David and Priestley, 1914). Within the last decade there appeared maximum and minimum reconstructions for late Wisconsin glacial conditions that attempted to explain available geological and glaciological data (Stuiver *et al.*, 1981, pp. 375–380). It is now apparent that both reconstructions are flawed (Denton *et al.*, 1988). For example, the CLIMAP maximum reconstruction (Stuiver *et al.*, 1981, p. 375) shows ice elevations that were too high in the southern Ross Embayment and over West Antarctica, where they conflict with interpretations drawn from the total gas content of the Byrd ice core (Raynaud and Whillans, 1982) and radio-echo layers in the West Antarctic Ice Sheet (Whillans, 1976). The CLIMAP minimum reconstruction, on the other extreme, is essentially the same as today's ice sheet in the Ross Embayment; this conflicts with glacial geologic evidence from Darwin/Hatherton and Beardmore Glaciers (Bockheim *et al.*, 1989; Denton *et al.*, 1989). Interim reconstruction designed to reconcile some of these problems (Drewry, 1979; Hughes *et al.*, 1985; Denton *et al.*, 1986, 1988) also have deficiencies (Denton *et al.*, 1988).

Figure 12 gives our new minimum (a) and maximum (b) reconstructions for the Ross ice drainage system based on the same Transantarctic outlet profiles but on two differing glaciological assumptions discussed earlier. Our chronology and reconstructions highlight the difference in behavior of the continental and marine portions of the Ross ice drainage system (Fig. 14). During late Wisconsin maximum glaciation, the Ross Embayment partially or wholly filled with grounded ice, while elevations of interior plateau ice remained nearly unchanged and perhaps even lowered slightly. During the late Wisconsin/Holocene deglacial hemicycle, grounded ice receded from the inner Ross Embayment while local Transantarctic glaciers expanded and the inland plateau surface changed very little and may even have increased slightly in surface elevation.

What mechanisms could have caused this behavior difference of marine and inland portions of the Ross ice drainage system? We argue from our glacial geologic data that the fundamental cause for the wide-

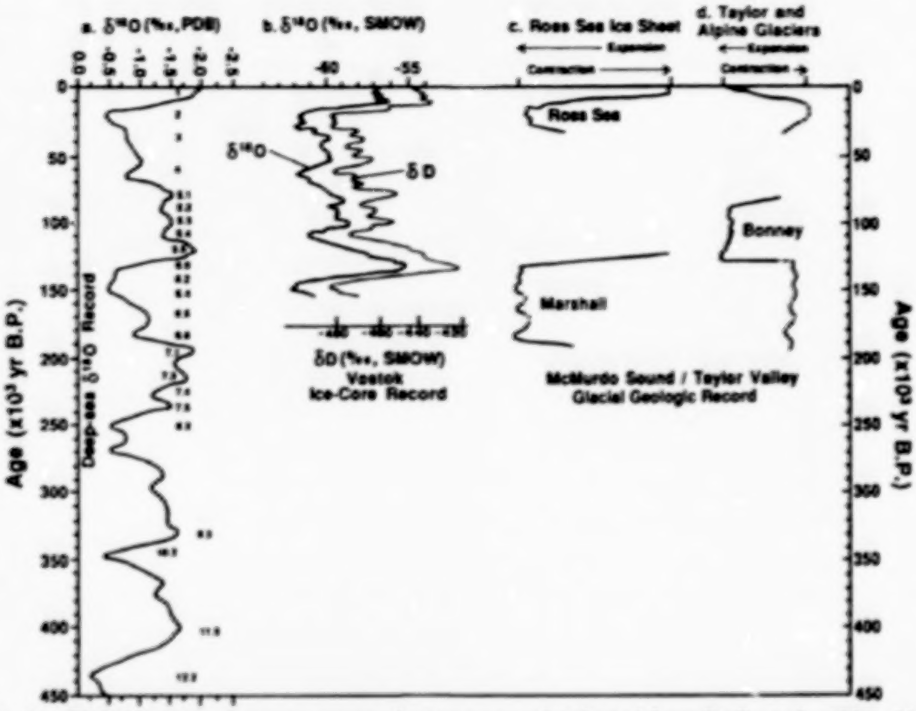


FIG. 14. Global average deep-sea oxygen-isotope record (adapted from Prell *et al.*, 1986), Vostok ice-core oxygen-isotope and deuterium records (adapted from Lorius *et al.*, 1985 and Jouzel *et al.*, 1987; chronology from ice model and accumulation rates), and McMurdo Sound/Taylor Valley glacial geology record (chronology from numerical dates). The record of alpine glaciers in the right-hand column includes Taylor Glacier, which drains the McMurdo Dome (Drewry, 1980, 1982).

spread grounding shown in Figures 12a and 12b must be sought in mechanisms that affected the Ross Embayment and not the inland plateau ice. We make this argument because, in view of decreased precipitation and little change in plateau elevations, it is unlikely that greater flow of inland ice through Transantarctic outlet glaciers caused grounding in the Ross Embayment. Therefore, such grounding must represent the response of the Ross Ice Shelf to reduced temperature, precipitation, and sea level. We agree with Hollin (1962) that sea-level variations, caused by Northern Hemisphere ice sheet fluctuations, are an important control on grounding-line variation and on pinning and unpinning the ice shelf. Grounding-line advance, accompanied by increased pinning of the ice shelf, could also cause northward migration of the calving front to the edge of the continental shelf, where the ice shelf could ground on relatively shallow submarine banks. This, in turn, could stimulate widespread grounding farther south. Ice-age temperature lowering may also aid ice-shelf grounding through decreased basal melting and increased protection of the calving front by pack ice. Both of these factors would operate solely in the Ross Embayment and could together overcome reduced precipitation to cause widespread grounding. We also argue from geologic evidence that the slight changes of the East Antarctic inland plateau and inland West Antarctica reflect the interplay between grounding and precipitation effects.

Changes of the marine and inland portions of the Ross ice drainage system correlate closely with the last and penultimate glacial/interglacial cycles (Fig. 14). Our identification of differing primary controls in these two portions of the drainage system is consistent with such behavior. The potential interdependence of these primary controls can explain the tight out-of-phase behavior that is so closely related to global glacial/interglacial cycles (Fig. 14). For example, low ice-age temperatures, which can enhance grounding in the Ross Embayment, are closely tied to reduced precipitation (Lorius *et al.*, 1985). Likewise, lower ice-age sea level promotes grounding in the Ross Sea, which removes a local moisture source and reduces inland penetration of storms.

The behavior of the Ross ice drainage system has global relevance to late Wisconsin sea level and climate. The maximum CLIMAP reconstruction illustrates thicker and more extensive grounded ice in the Ross Embayment and in West Antarctica (Stuiver *et al.*, 1981, Fig. 7–26) than we now show in Figures 12a or 12b. These two new reconstructions illustrate that the contribution of the Ross ice drainage system to sea-level lowering would depend on the relative magnitude of ice-volume loss (water equivalent) in the inland portion to volume gain (water equivalent above buoyancy) in the marine portion. Depending on the reconstruction eventually accepted for each portion, the sea-level contribution could be close to zero. If such a conclusion can be generalized, it means that the maximum CLIMAP reconstruction overestimated the Antarctic contribution to late Wisconsin sea-level fall (Denton and Hughes, 1981).

CONCLUSIONS

(i) From soil development, numerical dates, and glacial geologic data, we correlate Reedy III drift (Reedy Glacier), Beardmore drift (Beardmore Glacier), Britannia drifts (Hatherton and Darwin Glaciers), Ross Sea drift (McMurdo Sound/Taylor Valley), and "younger" drift (Terra Nova Bay). From 51 pertinent ^{14}C dates in Table 1 and an additional 15 from Clayton-Greene *et al.* (1988b), we conclude that Ross Sea drift is late Wisconsin (isotope Stage 2) in age.

(ii) From ice-surface profiles based on these drift sheets, we produce two reconstructions of the Ross ice drainage system at the height of late Wisconsin (isotope Stage 2) glaciation. Both show little elevation change of the interior polar plateau. The major differences occur in the extent of grounded ice in the Ross Embayment. One reconstruction shows limited grounded ice, whereas the other shows grounded ice to be extensive but thin. Both reconstructions show late Wisconsin ice-volume increase in the Ross ice drainage system to be considerably less than depicted by the maximum CLIMAP reconstruction (Denton and Hughes, 1981).

(iii) Radiocarbon dates from the western Ross Embayment show that late Wisconsin/early Holocene recession of grounded ice was underway by 13,040 yr B.P. and was complete by 6020–6600 yr B.P. Little change occurred in the East Antarctic plateau adjacent to the Transantarctic Mountains during massive recession of grounded ice from the Ross Embayment.

(iv) Sea-level variation and ice-shelf basal melting probably controlled grounding in the Ross Embayment. Precipitation variations and the grounding effect on Transantarctic Mountains outlet glaciers probably controlled minor elevation changes of the polar plateau adjacent to the Transantarctic Mountains.

(v) The Vostok and Dome Circe ice cores in the Ross ice drainage system are considered premier recorders of far-southern paleoclimate (Jouzel *et al.*, 1989). The Transantarctic Mountains outlet glacier profiles are consistent with this conclusion, in that they preclude an alternative explanation that the isotope records from these ice cores simply reflect extensive elevation changes of the East Antarctic polar plateau.

ACKNOWLEDGMENTS

The Antarctic field work and laboratory research at the University of Maine and the University of Wisconsin were funded by the Division of Polar Programs of the National Science Foundation. The Quaternary Isotopes Laboratory at the University of Washington is supported by the National Science Foundation. Logistic support in Antarctica was afforded by the National Science Foundation and the United States Navy. H. B. Conway, B. Hess, J. E. Leide, and R. Weed aided in the field work. T. J. Hughes gave valuable insights about ice-sheet reconstructions and lake-level fluctuations. This manuscript benefited greatly from reviews by T. J. Hughes, M. L. Prentice, J. F.

Splettstoesser, and D. E. Sugden. R. Kelly drafted the maps and longitudinal sections. N. Kealiher processed the manuscript.

REFERENCES

Alley, R. B., and Whillans, I. M. (1984). Response of the East Antarctic Ice Sheet to sea-level rise. *Journal of Geophysical Research* **89**, 6487–6493.

Bockheim, J. G. (1982). Properties of a chronosequence of ultraxerous soils in the Trans-Antarctic Mountains. *Geoderma* **28**, 239–255.

Bockheim, J. G., Wilson, S. C., Denton, G. H., Andersen, B. G., and Stuiver, M. (1989). Late Quaternary ice-surface fluctuations of Hatherton Glacier, Transantarctic Mountains. *Quaternary Research* **31**, 229–254.

Buntley, G. J., and Westin, F. C. (1965). A comparative study of developmental color in a Chestnut–Chernozem–Brunizem soil climosequence. *Soil Science Society of America Proceedings* **29**, 579–582.

Chinn, T. J. H. (1981). Hydrology and climate in the Ross Sea area. *Journal of the Royal Society of New Zealand* **11**, 373–386.

Clayton-Greene, J. M. (1986). "Proglacial sedimentation of late Wisconsin age in Miers Valley, Antarctica." M. Sci. thesis, University of Waikato, Hamilton.

Clayton-Greene, J. M., Hendy, C. H., and Denton, G. H. (1988a). The formation of debris mounds on Glacial Lake Miers. *Antarctic Journal of the United States*, in press.

Clayton-Greene, J. M., Hendy, C. H., and Hogg, A. G. (1988b). The chronology of a Wisconsin-aged proglacial lake in the Miers Valley, Antarctica. *New Zealand Journal of Geology and Geophysics*, in press.

Dagel, M. A., Hendy, C. H., Judd, F., Denton, G. H., and Stuiver, M. (1989). Stratigraphy and chronology of Stage 6 and 2 glacial deposits, Marshall Valley, Antarctica. *Boreas*, in press.

David, T. W. E., and Priestley, R. E. (1914). "Glaciology, Physiography, Stratigraphy, and Tectonic Geology of South Victoria Land: British Antarctic Expedition, 1907–1909. Vol. 1. Reports on the Scientific Investigations, Geology.

Debenham, F. (1921). "Recent and Local Deposits of McMurdo Sound Region: London, British Museum, British Antarctic (Terra Nova) Expedition (1910)." Vol 1, No. 3, pp. 63–90. Natural History Report. Geology.

Denton, G. H., and Armstrong, R. L. (1968). Glacial geology and chronology of the McMurdo Sound region. *Antarctic Journal of the United States* **3**, 99–101.

Denton, G. H., Armstrong, R. L., and Stuiver, M. (1970). Late Cenozoic glaciation in Antarctica. *Antarctic Journal of the United States* **5**, 15–22.

Denton, G. H., Armstrong, R. L., and Stuiver, M. (1971). The late Cenozoic glacial history of Antarctica. *In* "The Lake Cenozoic Glacial Ages" (K. K. Turekian, Ed.), pp. 267–306. Yale Univ. Press, New Haven, CT.

Denton, G. H., Bockheim, J. G., Wilson, S. C., Leide, J. E., and Andersen, B. G. (1989). Late Quaternary ice-surface fluctuations of Beardmore Glacier, Transantarctic Mountains. *Quaternary Research*, **31**, 183–209.

Denton, G. H., and Borns, H. W., Jr. (1974). Former grounded ice sheets in the Ross Sea. *Antarctic Journal of the United States* **9**, 167.

Denton, G. H., Borns, H. W., Jr., Grosswald, M. G., Stuiver, M., and Nichols, R. L. (1975). Glacial history of the Ross Sea. *Antarctic Journal of the United States* **10**, 160–164.

Denton, G. H., and Hughes, T. J. (1981). "The Last Great Ice Sheets." Wiley–Interscience, New York.

Denton, G. H., and Hughes, T. J. (1983). Milankovitch theory of ice ages, hypothesis of ice-sheet linkage between regional insolation and global climate. *Quaternary Research* **20**, 125–144.

Denton, G. H., Hughes, T. J., and Karlén, W. (1986). Global ice-sheet system interlocked by sea level. *Quaternary Research* **26**, 3–26.

Denton, G. H., Prentice, M. L., and Burckle, L. H. (1988). Cenozoic History of the Antarctic Ice Sheet. *In* "The Geology of Antarctica" (R. Tingey, Ed.). Oxford Univ. Press, London, in press.

Denton, G. H., Stuiver, M., and Austin, K. (1985). Radiocarbon chronology of the last glaciation in the McMurdo Sound region. *Antarctic Journal of the United States* **19**, 59–61.

Drewry, D. J. (1979). Late Wisconsin reconstruction for the Ross Sea region, Antarctica. *Journal of Glaciology* **24**, 231–244.

Drewry, D. J. (1980). Pleistocene biomodal response of Antarctic ice. *Nature (London)* **287**, 214–216.

Drewry, D. J. (1982). Ice flow, bedrock, and geothermal studies from radio-echo sounding inland of McMurdo Sound, Antarctica. *In* "Antarctic Geoscience" (C. Craddock, Ed.), pp. 977–983. Univ. of Wisconsin Press, Madison.

Drewry, D. J. (Ed). (1983). "Antarctica: Glaciological and Geophysical Folio." Scott Polar Research Institute, Cambridge.

Fillon, R. H. (1975). Late Cenozoic paleo-oceanography of the Ross Sea, Antarctica. *Geological Society of America Bulletin* **86**, 839–845.

Hendy, C. H., Healy, T. R., Rayner, E. M., Shaw, J., and Wilson, A. T. (1979). Late Pleistocene glacial chronology of the Taylor Valley, Antarctica, and the global climate. *Quaternary Research* **11**, 172–184.

Hollin, J. T. (1962). On the glacial history of Antarctica. *Journal of Glaciology* **4**, 173–195.

Hughes, T. J., Denton, G. H., and Fastook, J. L. (1985). Is the Antarctic Ice Sheet an analogue for Northern Hemisphere paleo-ice sheets? *In* "Models in Geomorphology" (J. J. Woldenburg, Ed.), pp. 25–72. Allen and Urwin, Boston.

Jouzel, J., Lorius, C., Petit, J. R., Genthon, C., Barkov, N. I., Kotlyakov, V. M., and Petrov, V. M. (1987). Vostok ice core: A continuous isotopic temperature record over the last climate cycle (160,000 years). *Nature* (*London*) 329, 403–408.

Jouzel, J., Raisbeck, G., Benoist, J. P., Yiou, F., Lorius, C., Raynaud, D., Petit, J. R., Barkov, N. I., Korotkevitch, Y. S., and Kotlyakov, V. M. (1989). A comparison of deep Antarctic ice cores and their implications for climate between 65,000 and 15,000 years ago. *Quaternary Research* 31, 135–150.

Judd, F. M. (1986). "The chronology of the Ross Sea II glaciation, an Antarctic glaciation of Illinoian age." M. Sci. thesis, University of Waikato, Hamilton.

Kellogg, T. B., Truesdale, R. S., and Osterman, L. E. (1979). Late Quaternary extent of the West Antarctic Ice Sheet: New evidence from Ross Sea cores. *Geology* 7, 249–253.

Lorius, C., Jouzel, J., Ritz, C., Merlivat, L., Barkov, N. I., Korotkevich, Y. S., and Kotlyakov, V. M. (1985). A 150,000 year climatic record from Antarctic ice. *Nature* (*London*) 316, 591–596.

Lorius, C., Raynaud, D., Petit, J. R., Jouzel, J., and Merlivat, L. (1984). Late glacial maximum–Holocene atmospheric and ice thickness changes from Antarctic ice core studies. *Annals of Glaciology* 5, 88–94.

Mabin, M. C. G. (1986). The Ross Sea section of the Antarctic ice sheet at 18,000 yr B.P.: Evidence from Holocene sea-level changes along the Victoria Land coast. *South African Journal of Science* 82, 506–508.

Mercer, J. H. (1968). Glacial geology of the Reedy Glacier area, Antarctica. *Geological Society of America Bulletin* 79, 471–486.

Mercer, J. H. (1972). Some observations on the glacial geology of the Beardmore Glacier area. *In* "Antarctic Geology and Geophysics" (R. J. Adie, Ed.), pp. 427–433. Universitetsforlaget, Oslo.

Nichols, R. L. (1968) Coastal geomorphology, McMurdo Sound, Antarctica. *Journal of Glaciology* 7 449–478.

Paterson, W. S. B., and Hammer, C. V. (1987). Ice core and other glaciological data. *In:* "North America and Adjacent Oceans During the Last Deglaciation" (W. F. Ruddiman and H. E. Wright, Jr., Eds.), pp. 91–109. Geological Society of America, Boulder, CO.

Péwé, T. L. (1960). Multiple glaciation in the McMurdo Sound region, Antarctica, a progress report. *Journal of Geology* 68, 489–514.

Prell, W. L., Imbrie, J., Martinson, D. G., Morley, J. J., Pisias, N. G., Shackleton, N. J., and Streeter, H. F. (1986). Graphic correlation of oxygen isotope stratigraphy: Application to the Late Quaternary. *Paleoceanography* 1, 137–162.

Raynaud, D., and Whillans, I. M. (1982). Air content of the Byrd core and past changes in the West Antarctic Ice Sheet. *Annals of Glaciology* 3, 269–273.

Ross, Captain Sir James Clark (1847). "A Voyage of Discovery and Research in the Southern and Antarctic Regions During the Years 1839–1843," Vols. 1 and 2. John Murray, London.

Scott, R. F. (1905). "The Voyage of the Discovery," Vols. 1 and 2. Scribner's, New York.

Stuiver, M., Denton, G. H., Hughes, T. J., and Fastook, J. L. (1981). History of the marine ice sheet in West Antarctica during the last glaciation, a working hypothesis. *In* "The Last Great Ice Sheets" (G. H. Denton and T. H. Hughes, Eds.), pp. 319–436. Wiley–Interscience, New York.

Stuiver, M., and Polach, H. S. (1977). Discussion-reporting of ^{14}C data. *Radiocarbon* 19, 355–363.

Whillans, I. M. (1976). Radio-echo layers and the recent stability of the West Antarctic Ice Sheet. *Nature* (*London*) 264, 152–155.

MARINE RECORD OF LATE QUATERNARY GLACIAL-INTERGLACIAL FLUCTUATIONS IN THE ROSS SEA AND EVIDENCE FOR RAPID, EPISODIC SEA LEVEL CHANGE DUE TO MARINE ICE SHEET COLLAPSE

John B. Anderson
Department of Geology and Geophysics
Rice University
Houston, Texas, 77251

INTRODUCTION

Some of the questions to be addressed by SeaRISE include: 1) what was the configuration of the West Antarctic ice sheet during the last glacial maximum; 2) what is its configuration during a glacial minimum; and 3) has it, or any marine ice sheet, undergone episodic rapid mass wasting? This paper addresses these question in terms of what we know about the history of the marine ice sheet, specifically in Ross Sea, and what further studies are required to resolve these problems. A second question concerns the extent to which disintegration of marine ice sheets may result in rises in sea level that are episodic in nature and extremely rapid, as suggested by several glaciologists (Clark and Lingle, 1977; Thomas and Bentley, 1978; Hughes, 1987). Evidence that rapid, episodic sea-level changes have occurred during the Holocene is also reviewed.

LATE WISCONSIN-HOLOCENE GLACIAL HISTORY OF THE ROSS SEA

Scott (1905) was the first to speculate that Ross Sea once was covered by an extensive ice sheet, which he felt was grounded on the continental shelf as far north as Cape Adare (Fig. 1). He called this expanded ice sheet the "Great Ice Sheet." David and Priestly (1914), who corroborated Scott's ideas, discovered foreign erratics and moraines perched well above present sea level (up to 305 m) at Cape Royds and Ross Island (Fig. 1). From this they concluded that the "Great Ice Sheet" once extended approximately 320 km north of its present position. Later, Priestley (1923) discovered metamorphic and dioritic erratics resting on the basaltic rocks of Cape Adare. He inferred that the East Antarctic Ice Sheet (EAIS) once had flowed over and through the mountains of northern Victoria Land and onto the adjacent continental shelf, filling Robertson Bay (Fig. 1) with at least 300 m of ice. Stuiver et al. (1981) cite widespread occurrences of striated surfaces, erratics, and moraines, now located well above the present ice surface, as evidence for former expansion of the ice sheet in the Ross Sea region. The most widespread deposit is the "Ross Sea Drift," perched 240 m to 610 m above present sea level.

Numerous radiocarbon dates obtained in the Ross Sea region indicate that the late Wisconsin glacial maximum occurred between 21,200 and 17,000 yrs. B.P. (Stuiver et al., 1981). During the glacial maximum, approximately 1,325 m of ice is believed to have been grounded in McMurdo Sound (Stuiver et al., 1981). This ice sheet pushed westward into the Dry Valleys, and grounded in Ross Sea to the edge of the continental shelf (Stuiver et al., 1981). Perched erratics, some resting at elevations of 320 m above present sea level, on Beaufort and Franklin Islands (Fig. 1), provide evidence that an ice sheet once extended onto the continental shelf. Alternatively, these erratics could represent ice-rafted deposits, placed prior to the uplift of the islands, which occurred after the initiation of ice-rafting in Ross Sea (Stuiver et al., 1981). Based on the continental record, the ice front began its retreat around 17,000 yrs. B.P. and reached its present position by 6,190 yrs. B.P. (Stuiver et al., 1981).

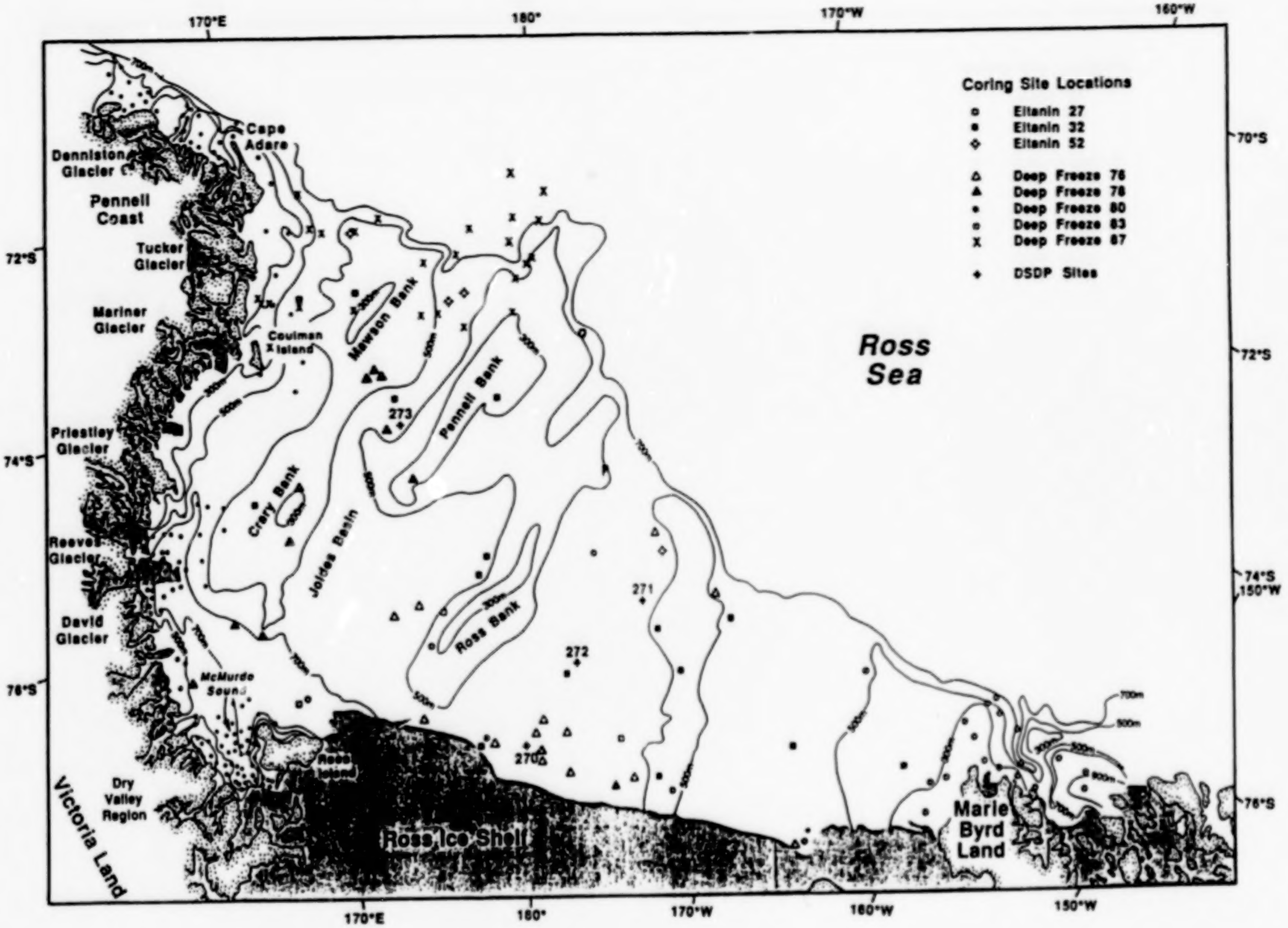


Figure 1. Geography and bathymetry of the Ross Sea. Also shown are locations of piston cores that have been examined in detail.

Arguments for a more extensive ice sheet in Ross Sea during the last glacial maximum have not gone uncontested. As early as 1921, Debenham argued that localized glacial expansions could explain the presence of raised moraines and erratics in the southern Victoria Land region. Mayewski (1975) mapped ancient moraines associated with Scott, Amundsen, Shackleton and Beardmore glaciers and concluded that they were deposited by a much thickened EAIS and that the grounding line of Ross Ice Shelf could have been located very near its present position during these events. Also, Whillans (1976) argued that the West Antarctic Ice Sheet (WAIS) probably varied little in size and dimensions during the last 30,000 years, based on glaciological arguments.

Drewry (1979) reviewed the evidence for ice sheet expansion onto the Ross Sea continental shelf during the last glacial maximum. He concluded that the grounding line of the ice sheet occurred only slightly seaward of its present position 18,000 years ago, but that the ice shelf might have extended over much of the continental shelf. His argument draws from the concept that a eustatic sea-level fall of greater than 120 to 130 m sustained over a period of 5,000 to 10,000 years must occur to fully ground the ice sheet. It is estimated that the maximum sea-level fall around Antarctica, caused by the late Wisconsin build-up of ice in the northern hemisphere, was only on the order of 75 to 100 meters (Lingle and Clark, 1979).

Marine Geological and Geophysical Evidence for an Expanded Ice Sheet

Results of Seismic Investigations

Listed below are some of the questions that high resolution seismic data allow us to address.

1) Are there geomorphological features on the sea floor which mark former grounding lines for the last glacial maximum? If so, how many grounding lines can be identified by this method and what do they imply about geographic, bathymetric and geological controls on the grounding line position?
2) Did the positions of the major ice streams shift significantly during the last few glacial cycles?
3) Are there geomorphological features on the shelf which provide clues to the dynamics and stability of the grounding line (i.e., the subglacial deltas of Alley et al., 1989)?

More miles of seismic reflection tracklines exist for Ross Sea than any other part of Antarctica. Data from earlier surveys consist mainly of sparker profiles, which are short and seldom tie (Fig. 2). More recent seismic surveys concentrate on examining the deep structure and stratigraphy of the region and utilized large air guns as the sound source. The stratigraphic resolution of this method is poor, so these data have limited value in addressing the problems outlined by the SeaRISE workshop participants.

The most recent seismic survey of the Ross Sea continental shelf was conducted during USAP 90 onboard the R/V Polar Duke (Fig. 3). Data acquisition employed a new sound source, a bubble-free air gun. These data provide greater stratigraphic resolution and greater aerial coverage than previously available for the region. Still, the resolution of these data falls short of that needed to address some of those questions related to the SeaRISE program.

A series of north-northeast to south-southeast oriented basins and ridges characterizes the bathymetry of Ross Sea (Fig. 1). These features extend beneath Ross Ice Shelf where they display a more northwest to southeast orientation (Drewry, 1983). Hughes (1977) observed that the linear basins of the Ross Sea shelf roughly correspond to the locations of modern ice streams at the grounding line of Ross Ice Shelf. Based on this observation, he argued that the linear basins of the shelf formed by accelerated glacial erosion beneath ice streams during a previous expansion of the WAIS. However, Vanney et al. (1981) argued that the basin and ridge topography of the Ross Sea shelf also corresponds to the tectonic fabric of the shelf, hence these features may not necessarily

BLANK

PAGE

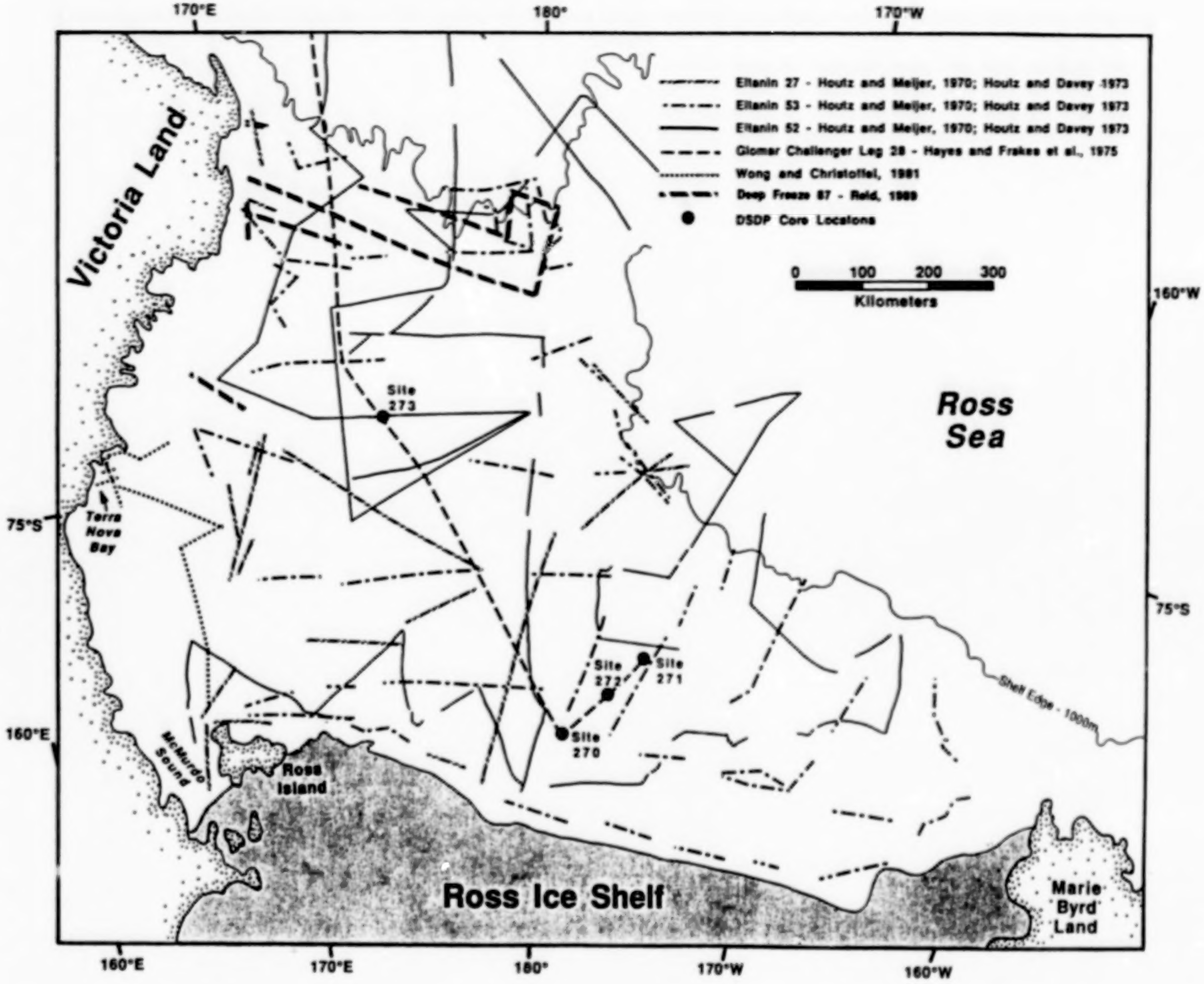

170°E
180°
170°W
160°W
75°S
160°E
170°E
180°
170°W
160°W
75°S
Eltanin 27 - Houtz and Meijer, 1970; Houtz and Davey 1973
Eltanin 53 - Houtz and Meijer, 1970; Houtz and Davey 1973
Eltanin 52 - Houtz and Meijer, 1970; Houtz and Davey 1973
Glomar Challenger Leg 28 - Hayes and Frakes et al., 1975
Wong and Christoffel, 1981
Deep Freeze 87 - Reid, 1989
DSDP Core Locations
0 100 200 300
Kilometers
Victoria Land
Ross Sea
Site 273
Site 272
Site 271
Site 270
Terra Nova Bay
McMurdo Sound
Ross Island
Shelf Edge - 1000m
Ross Ice Shelf
Marie Byrd Land

Figure 2. High resolution seismic tracklines for the Ross Sea (prior to 1990).

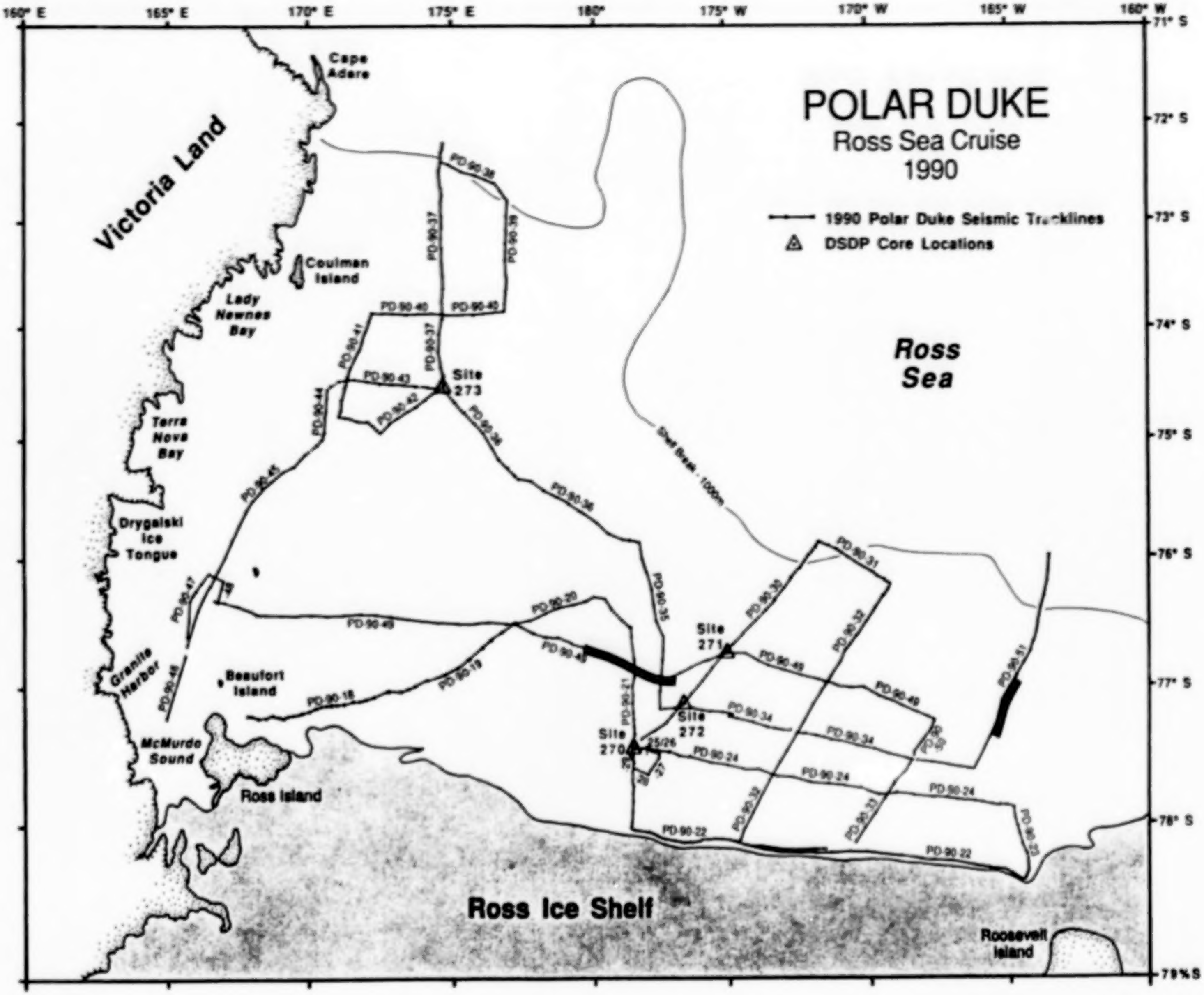


Figure 3. Polar Duke-USAP 90 seismic tracklines. These data were acquired using a new, high resolution seismic source, a bubble-free air gun. Highlighted portions of the tracklines are illustrated in the text.

be of glacial origin. Alternatively, we might infer that there is a geological control on the positions of ice streams.

During USAP 90, several east-west profiles were acquired across the shelf to address whether there is, or has been, a geological control on the position of ice steams in the Ross Sea region (Lines PD90-18, 22, 24, and 49, Fig. 3). These profiles show that large-scale cut-and-fill structures characterize the upper stratigraphic sequence (Fig. 4). The width to depth ratio of these scours mimics that of the modern ridge and basin topography, hence their interpretation as relict glacial troughs. The locations of these features on the shelf has varied through time and tectonic controls on trough positions is observed only in the western Ross Sea. Thus, modern troughs apparently mark the positions of former ice streams, as suggested by Hughes (1977), but the locations of these ice streams have changed with time and are not controlled by geological features on the shelf, at least not on the exposed portions of the shelf.

Early studies of seismic reflection records from the continental shelf led to the discovery of a widespread unconformity, the Ross Sea Unconformity (RSU), inferred to be a glacial erosional surface (Houtz and Davey, 1973; Karl et al., 1987). Efforts to date this unconformity yielded a range of dates (Kellogg et al., 1979, [18000 B.P.]; Savage and Ciesielski, 1983 [0.65 to 13.8 Ma]; Hayes and Frakes, 1975 [3 to 5 Ma]). A more recent examination of these same records led to the conclusion that the RSU represents an amalgamation of several erosional surfaces that cuts across different stratigraphic levels (Bartek, 1989), a conclusion supported by USAP 90 seismic records.

On the inner-most shelf, the RSU is a sea floor unconformity where acoustic basement outcrops all along the coast. There, conspicuous evidence for glacial erosion of the sea floor exists as glacial troughs (Shepard, 1931). These broad, deep submarine valleys, located offshore of outlet glaciers and ice streams, are U-shaped and foredeepened. The largest of these is Drygalski Trough (Fig. 5). This feature exists offshore of David Glacier where ice from East Antarctica flows through the Transantarctic Mountains. Seismic records across the trough provide evidence for several episodes of glacial erosion (Fig. 5).

Seismic reflection profiles also provide evidence of subglacial deposition on the shelf. Wong and Christoffel (1981) mapped one of the more spectacular examples of a morainal bank along the northern rim of Drygalski Trough. It is comprised of steeply dipping (to the north) reflectors which abruptly terminate at the southern flank of the trough (Fig. 5) and rests above the RSU. They interpreted this feature as a "delta moraine," and suggested that it marks the northern limit of Ross Ice Shelf during a previous glacial advance.

USGS profile 413 from the western Ross Sea (Karl et al., 1987) shows a wedge-shaped body of deposits which thickens to the north and has seaward dipping clinoforms at its northern end. Internal reflectors are discontinuous and hyperbolic reflectors characterize the upper surface (Fig. 6a). Similar features, recognized in seismic records acquired during Deep Freeze 87 and USAP 90 (Fig. 4), are identical to acoustically massive wedges described in high resolution seismic records from the southeastern Canadian shelf and interpreted as till tongues (King and Fader, 1986).

USAP 90 seismic profiles on the central and eastern portions of the continental shelf display several mound-like features with foreset beds (Fig. 6b). These features, interpreted tentatively as subglacial deltas (Bartek and Anderson, in press, a), possibly were produced by processes similar to those acting beneath Ice Stream B today (Alley et al., 1989). These features, best seen in seismic lines acquired perpendicular to the paleodrainage divides, rest on glacial erosional surfaces.

High resolution seismic records from the western Ross Sea (Deep Freeze 87) show a change in the character of near surface deposits from north to south within this area (Reid, 1989). The southern part of the area is characterized by intercutting erosional surfaces (former troughs), relatively high relief on individual reflectors, till tongues and morainal banks (Fig. 5).

BLANK
PAGE

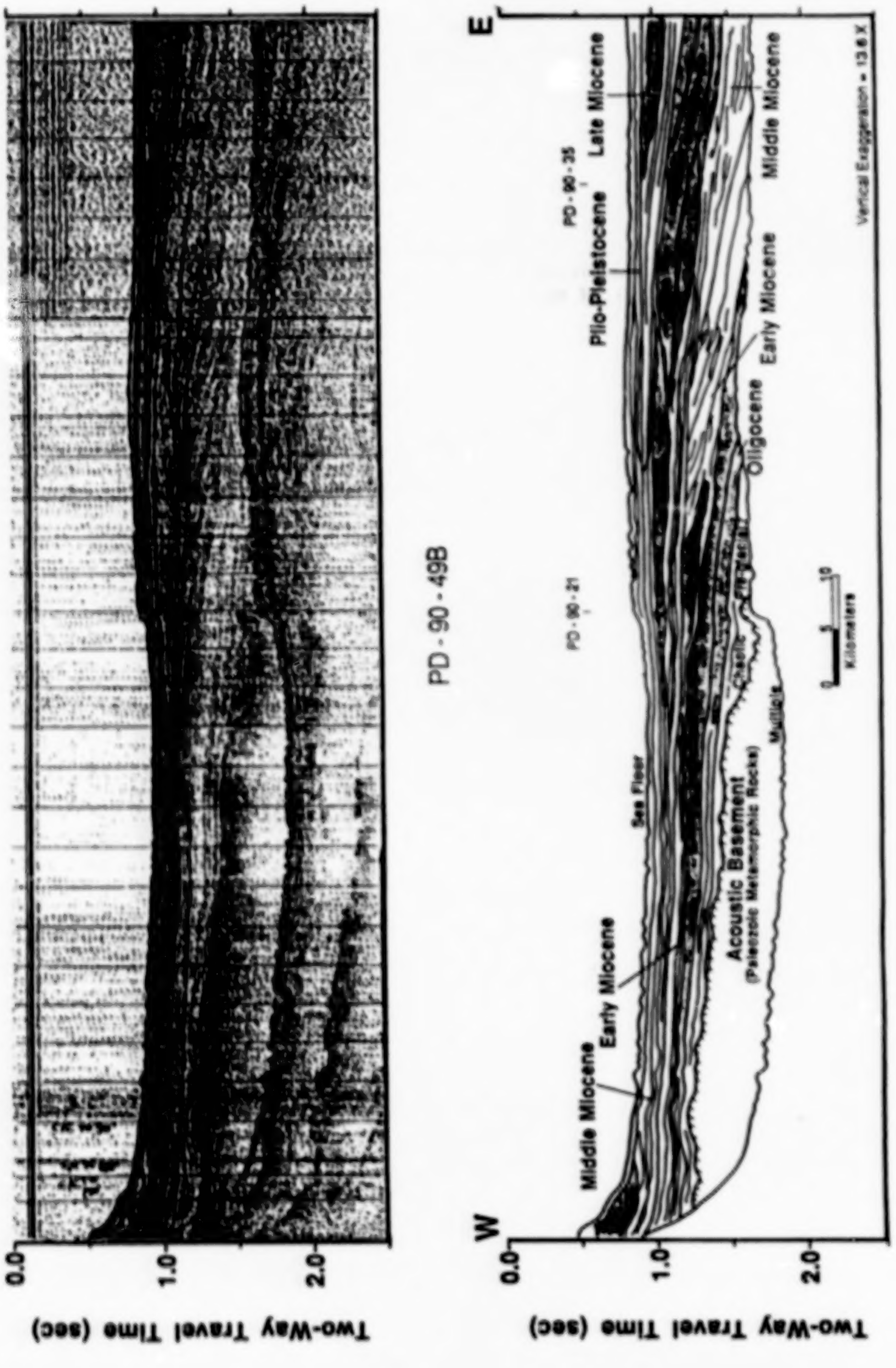


Figure 4. Polar Duke-USAP 90 seismic line PD-90-49B showing broad-scale erosional surfaces, glacial troughs, and massive sediment bodies that are interpreted as "till tongues". See figure 3 for profile location.

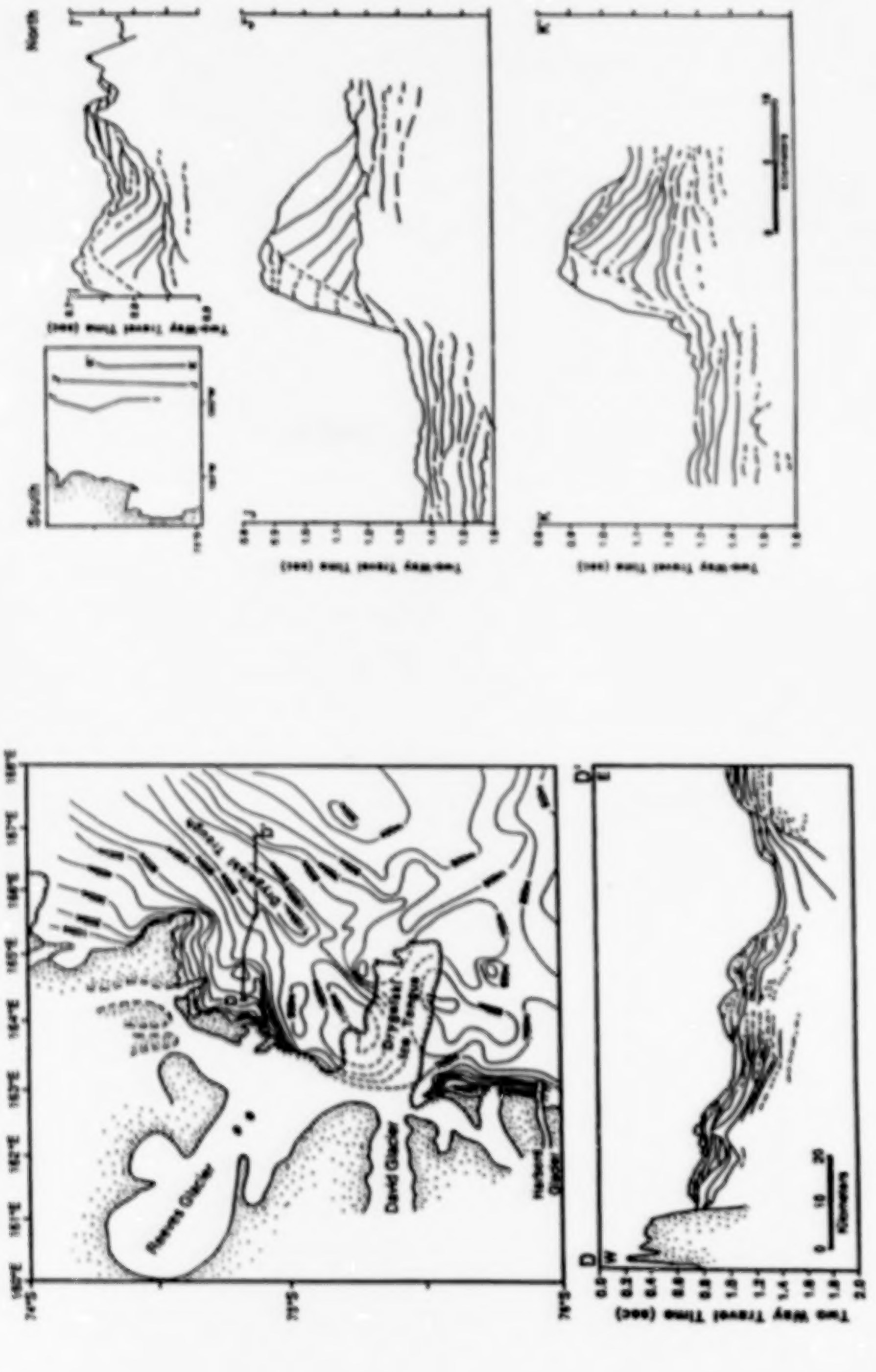


Figure 5. Bathymetric map of Drygalski Trough. Seismic records from this feature provide evidence for several episodes of glacial erosion, as shown in profile D-D". A large "morainal feature that rests above the upper-most unconformity is shown in interpreted seismic lines (I', J', and K') from the northwestern flank of the trough. Seismic lines from Wong and Christoffel, 1981.

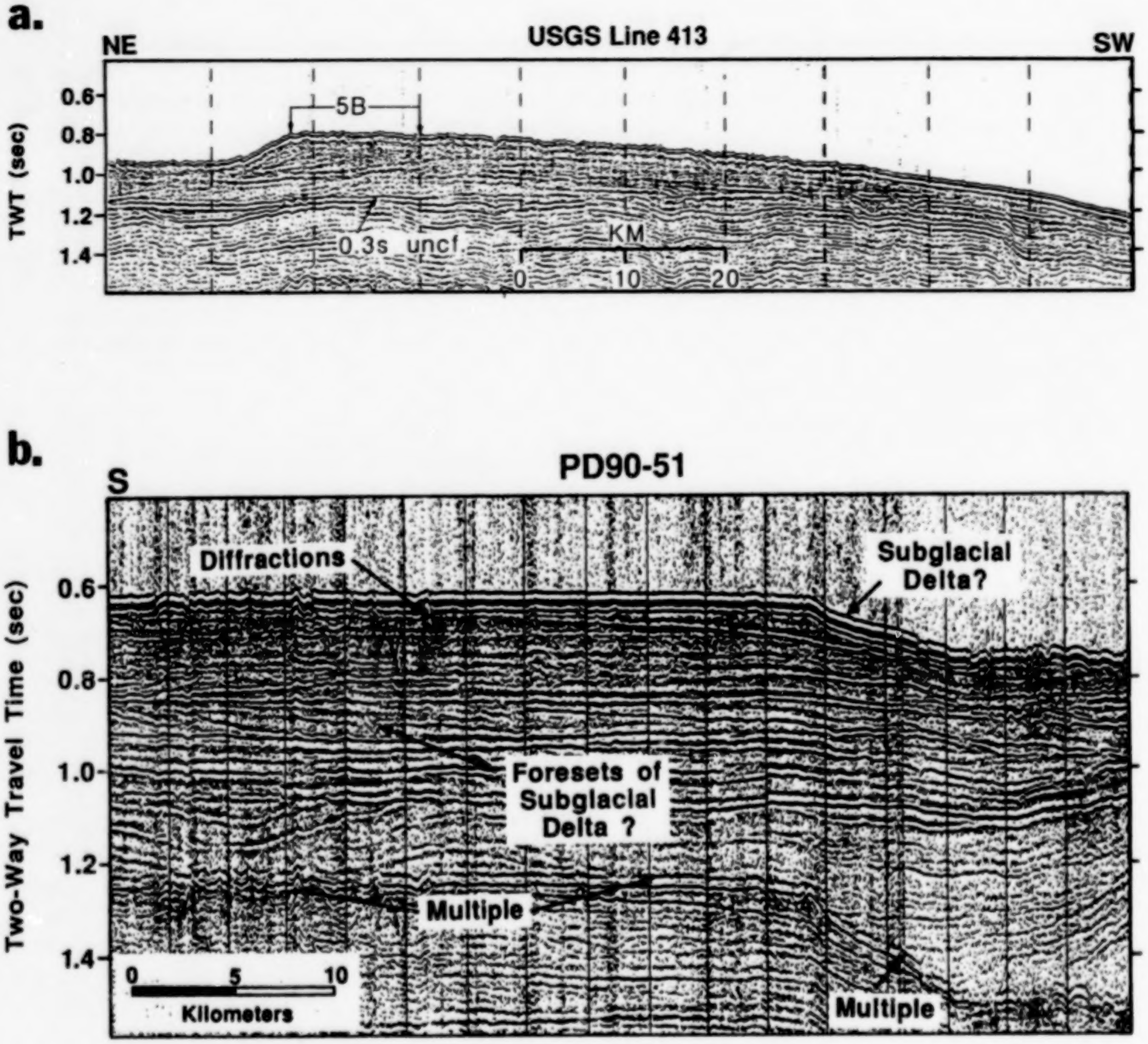


Figure 6. A. USGS seismic profile 413 from the western Ross Sea showing wedge-shaped body (from Karl et al., 1987). This feature is similar to "till tongues" of the southeastern Canadian shelf as described by King and Fader (1986). B. USAP 90 seismic line PD90-51 showing a similar wedge-shaped body, interpreted as a possible subglacial delta. See figure 3 for profile location.

The northern part of the survey area shows generally flat, coherent reflectors separated by acoustically transparent zones. Relief on individual reflectors indicates a sea floor topography that is relatively broad and gentle, and without glacial troughs. Reid (1989) suggests that the boundary between these different seismic provinces marks the northern limit of an ice sheet grounding line during the late Wisconsin.

In summary, high resolution seismic reflection profiles from Ross Sea provide abundant evidence that ice sheets were grounded there on more than one occasion. However, most of these seismic data consist of sparker and air gun records, and the stratigraphic resolution obtained with these seismic sources is not adequate to address some of the questions raised in the SeaRISE workshop.

To address these questions, we must undertake systematic high resolution seismic surveys in which track lines are planned using available information about paleodrainage of the late Wisconsin ice sheet (discussed in the following section). These surveys should employ high frequency sound sources, such as small water guns and/or a Huntec system, in conjunction with side-scan sonar mapping. Holocene sediments on the Ross Sea continental shelf are relatively thin, therefore, subglacial features with greater than approximately 1 m relief will be detectable using side-scan sonar images.

Results of Sedimentologic and Petrographic Studies

Many piston cores have been collected from the Ross Sea continental shelf (Fig. 1). A number of investigators have examined these cores in an attempt to reconstruct the late Quaternary glacial setting (Fillon, 1975; Kellogg and Truesdale, 1979; Kellogg et al., 1979; Anderson et al., 1980, 1984; Domack et al., 1980; Edwards et al., 1987; Reid, 1989) . In general, the late Pleistocene-Holocene stratigraphy of the shelf, revealed by these cores, consists of Holocene glacial marine sediments, reflecting a dominance of marine influence on sedimentation (Chriss and Frakes, 1972), resting on, and typically in sharp contact with, diamictons. These diamictons reflect the earlier influence of glacial processes on sedimentation (Kellogg et al., 1979; Anderson et al., 1980; 1984). A key question concerns the subglacial versus glacial marine origin of these diamictons. Fillon (1975) favored a glacial marine origin for these deposits, whereas Kellogg et al. (1979), Anderson et al. (1980; 1984), and Domack et al. (1980) argued in favor of a subglacial origin.

The evidence provided by Kellogg et al. (1979) for a subglacial origin of diamictons (their unit D) includes: 1) the widespread occurrence of this unit; 2) its compact nature and lack of sorting; and 3) the presence of reworked microfossils. Anderson et al. (1980; 1984) and Domack et al. (1980) re-examined these diamictons and added to the list of glacial criteria textural and mineralogic homogeneity of individual units, pebble fabric (relative to the vertical plane), and pebble shape. These combined criteria imply rapid deposition from basal debris zones, but do not allow us to conclude whether these diamictons were the products of subglacial sedimentation (basal tills) or resulted from deposition beneath an expanded ice shelf. Diamictons produced by sedimentation beneath ice shelves and by sediment mass movement are quite similar to those deposited by grounded ice (Kurtz and Anderson, 1979; Anderson et al., 1980, in press).

Diamictons from the Ross Sea shelf typically have shear strength values that exceed 2.5 kg/cm^2 (Anderson et al., 1980). Overcompaction implies subglacial deposition, but also can result from sediment overburden. Edwards et al. (1987) performed geotechnical tests on these diamictons and concluded that they were overcompacted, but the degree of overcompaction is minor and implies an ice sheet thickness only tens of meters greater than the water depth of the shelf. However, the piston cores Edwards and his colleagues examined penetrated only a short distance into the diamicton. Hence, they measured the physical properties of sediments deposited just prior to when the ice sheet decoupled from the sea floor.

One of the most crucial criteria for determining a subglacial and/or sub-ice shelf origin of diamictons involves petrographic analysis. Since the turn of the century, glacial geologist have utilized petrographic data (mineral and clast content of glacial deposits) to identify tills and determine the source area of these deposits. These studies allowed them to construct glacial paleodrainage maps. Basal tills and sub-ice shelf deposits should consist of fewer varieties of rocks and minerals than iceberg-rafted sediments, whose source area may be quite extensive. The provinance of the tills can be established where the geology of the source area is known. Domack (1982) demonstrated that glacial sediments from the continental shelf off Wilkes Land correlate positively with exposures of continental rocks. Results of diamicton studies on the Weddell Sea continental shelf (Anderson et al., in press), and in Marguerite Bay (Kennedy and Anderson, 1989) parallel the findings off Wilkes Land.

Anderson et al. (1984; submitted) summarize the results of petrographic analyses performed by Balshaw (1981), Myers, (1982), and Reid (1989) on four components of Ross Sea diamictons (rock clasts, coarse sands, heavy minerals, and clay minerals) and the glacial marine sediments that overlie them. Factor analysis was used to group diamictons of similar composition. When plotted on a map, the results of these petrographic analyses fall into seven provinces whose boundaries follow the northeast-southwest bathymetric trend of the shelf and provide the basis for a paleodrainage reconstruction for the late Wisconsinan ice sheet (Fig. 7). Truswell and Drewry (1984) examined reworked palynomorphs in diamictons from the central and eastern Ross Sea shelf and found distribution patterns broadly similar to our petrographic provinces. The paleodrainage map shown in Figure 7 also shows broad parallels with the paleodrainage map of Denton and Hughes (1981), however, the author hastens to emphasize that the actual number of cores examined from the eastern and central portions of the shelf was relatively small, and additional work is needed to better constrain the paleodrainage divides shown in Figure 7.

The most thorough petrographic and sedimentological studies of piston cores have been conducted in the western Ross Sea (Myers, 1982; Reid, 1989) and in McMurdo Sound (Bartek, 1989). Half (20) of the cores collected from the western Ross Sea shelf penetrated diamictons that bear all the characteristics of basal till. The other twenty cores penetrated either glacial marine sediments or sediment gravity flows. The latter deposits appear to thicken to the north and toward the coast and may overly basal tills in these areas. North of Coulman Island (Fig. 1) piston cores penetrated a wide range of glacial marine sediments, including carbonates, but no overcompacted diamictons. This suggests that the late Wisconsinan grounding line at one time existed at the approximate latitude of Coulman Island (Fig. 7, Reid, 1989). Diamictons from the southwestern Ross Sea separate into three petrographic provinces whose boundaries correspond to the northeast-southwest-oriented bathymetric features on the shelf (Fig. 7).

Piston cores from McMurdo Sound penetrated glacial marine sediments, diatomaceous muds, and a variety of sediment gravity flow deposits; the latter group consists of sediments derived from both Ross Island and the Dry Valleys (Bartek, 1989; Bartek and Anderson, in press, b). Basal tills were not collected, although absence of a lag deposit associated with unconformities penetrated in piston cores indicates that glacial ice caused erosion, instead of marine currents. Thus, piston cores from McMurdo Sound indicate a recent period of erosion followed by sedimentation of glacial marine sediments, with a large biogenic component (diatom frustules), and sediment gravity flow deposits. Anderson et al. (1984) suggest that the absence of tills in McMurdo Sound may imply that grounded, stagnant ice existed there during the late Wisconsin.

Retreat of Ice Sheet from Continental Shelf

Reid (1989) studied piston cores collected north and south of Coulman Island for purposes of reconstructing the glacial setting during and after the glacial maximum. He obtained radiocarbon dates between >35,510 B.P. and 17,390 +/- 500 B.P. for glacial marine deposits recovered in piston cores north of the island. Examination of these deposits and correlation with

BLANK
PAGE

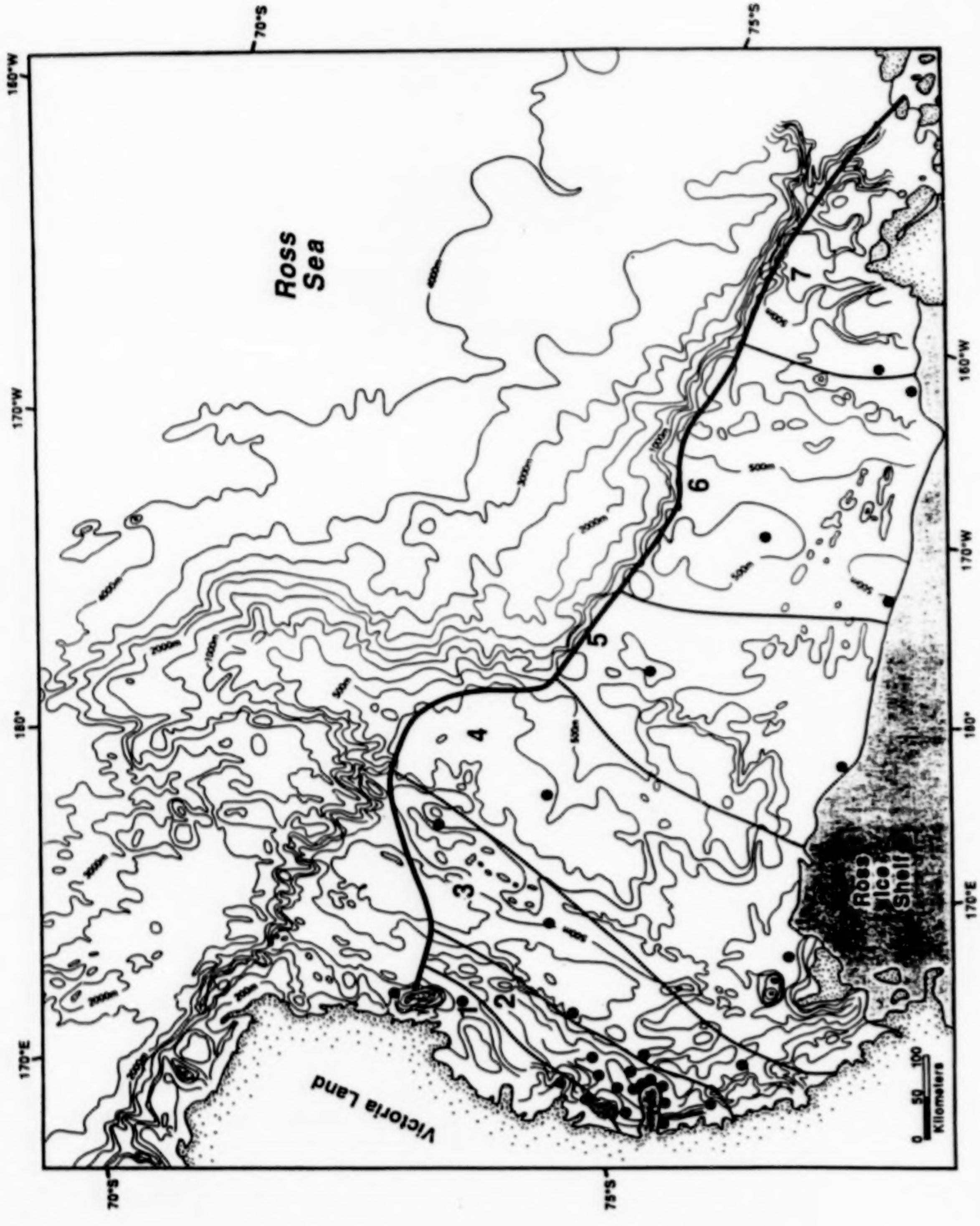


Figure 7. Petrographic provinces for Late Wisconsanan basal tills of Ross Sea and reconstructed ice sheet grounding line.

cores acquired south of the island led to the conclusion that sediment deposition occurred seaward of a grounded ice sheet and that icebergs, not subglacial ice, delivered the sand and pebble-sized glacial debris to these sites. From this, he deduced that either an ice shelf was not present at the outer shelf, or the ice shelf retreated from this area rapidly. The presence of warm deep water currents impinging onto the shelf, similar to the situation today (Jacobs et al., 1985), may account for the absence of an ice shelf at the outer shelf.

The vast majority of piston cores from the central and eastern Ross Sea contain diamictons directly overlain by diatomaceous muds and oozes. A sharp contact separates these units, indicating that the change from subglacial sedimentation to hemipelagic sedimentation occurred rapidly. Grain size analysis of surface sediments reveals a strong pattern of lateral size grading in which sediments decrease in grain size and become more diatomaceous to the south. This may be due to an increase in water depth and a decrease in marine current energy in that direction, and reflects the present dominance of marine processes on sedimentation (Anderson et al., 1984; Dunbar et al., 1985).

Ledford-Hoffman et al. (1986) obtained Pb^{210} sedimentation rates for diatomaceous sediments overlying diamictons in the southwestern Ross Sea, south of 75° S. These rates vary from less than 1 mm/y to 2.7 mm/y, and the thickness of these deposits varies from less than 10 cm to over two meters. If pelagic sedimentation began soon after the ice sheet withdrew from this portion of the shelf, and if these rates have remained relatively constant, the ice sheet retreated only a few thousand years B.P. These combined results suggest that the grounding line retreated rather slowly from north to south, or that it stabilized for some time at a position somewhere between Coulman Island and 75° S.

Leventer and Dunbar (in press) studied the more recent glacial and sea ice history of the McMurdo Sound region. They recognize a pattern in which the extent and/or duration of coastal polynyas increased during the Little Ice Age. They attribute this pattern to colder air temperatures and more persistent winds.

In summary, tills recovered in piston cores from the Ross Sea continental shelf indicate widespread grounding of the ice sheet during the last glacial maximum. Core coverage in the eastern and central Ross Sea is too sparse to allow detailed reconstruction of grounding line positions and paleodrainage on the shelf during the last glacial maximum. Core coverage is better in the western Ross Sea and the glacial setting during the late Wisconsin is better known, but the history of ice sheet retreat from the shelf remains problematic. What is needed is the acquisition of core transects extending from the front of Ross Ice Shelf to the edge of the continental shelf. An additional transect should extend east-west across the shelf to allow the reconstruction of paleodrainage in the region. These cores, if acquired in conjunction with high resolution seismic records (3.5 KHz and small water gun or Huntec) should assure stratigraphic integrity; USAP 90 seismic records from the region show several unconformities in the upper sediment column. In addition, side-scan sonar profiles may provide a direct record of subglacial features and can be used to avoid sampling areas subject to intense iceberg scour.

Detailed radiometric work may help to constrain the timing of ice sheet retreat from the shelf; this requires age control on the glacial marine sediments directly overlying tills. Reliable radiometric dates have been obtained for diatomaceous sediments of the Wilkes Land continental shelf (Domack et al., 1989) and for Prydz Bay (Domack and Jull, in press) using the tandem accelerator method.

DEFINING THE CONDITIONS DURING AN EXTREME GLACIAL MINIMUM

One question that remains unresolved in the late Quaternary glacial history of Ross Sea concerns the extent of grounded and floating ice over the continental shelf during a glacial minimum. This is an important problem; it bears on whether the ice sheet and ice shelf could experience further retreat before reaching a stable configuration, as suggested by Hughes (1973).

The global isotopic and sea-level record provides little information with regard to this question. The oxygen isotope record for the world ocean provides evidence that global temperatures and/or ice volume during Pleistocene interglacials were close to, and potentially less than, those of the present (Imbrie et al., 1984). Sea-level curves for the late Pleistocene indicate that sea level approximately 120,000 yrs. B.P. was a few meters higher than that of the present day (Kendall and Lerche, 1988).

The geological approach to this problem involves drilling through the ice shelf and searching for evidence that open marine conditions existed there during previous interglacials. This is not a "fool-proof" method; the ice sheet undoubtedly has removed older Pleistocene deposits during the late Wisconsin advance. The sediments recovered at RISP J 9 include abundant diatoms, implying open marine conditions. However, the age of these sediments remains the subject of considerable debate. One group of scientists contends that there was no recovery of Plio-Pleistocene sediments (Harwood et al., 1989) while the other group argues that Pleistocene diatoms were sampled and their presence implies open marine conditions at this site, perhaps as recently as 18,000 to 20,000 yrs. B.P. (Kellogg and Kellogg, 1986). It remains unclear whether paleontological resolution is at a stage where Pleistocene deposits will be recognized if they are recovered. Clearly, more work must be undertaken to resolve this problem.

NATURE OF THE HOLOCENE SEA LEVEL RISE

In his summary of Holocene sea-level curves, Kidson (1982) describes two schools of thought which concern the nature of sea-level rise since the late Wisconsinan glacial maximum. One, which he calls the Shepard school, includes those who favor a more or less continuous rise in sea level since about 18,000 B.P.. The other, the Fairbridge school, includes those workers who favor a more episodic post glacial rise in sea level. To some degree this controversy stems from a lack of prior knowledge of hydro-isostatic and geoidal influences, which results in non-uniform changes in sea-level in different parts of the world (Peltier, 1980). Some early attempts to establish sea level curves using data gathered from different portions of the globe (i.e. Fairbridge, 1961) resulted in step-like curves (Fig. 8) that may simply be an artifact of these effects (Kidson, 1982). On the other hand, some curves for specific regions, such as those from the New Zealand and Australian shelves (Carter et al., 1986) and the Gulf of Mexico shelf (Fig. 9), provide evidence of episodic rises in sea level.

In part, the episodic sea-level school of thought has suffered from the lack of a mechanism that could explain the rapid rises indicated by some curves. For example, data from the Gulf of Mexico (Rehkemper, 1969; Nelson and Bray, 1970; Frazier, 1974, Fig. 9) indicate rapid rises in sea level of as much as 3 to 5 cm/yr. Mass wasting of marine ice sheets is the most plausible mechanism for such rapid changes.

Geochemical and Stratigraphic Proxies for Examining Deglaciation Mechanisms

The stratigraphy of continental shelves and the geochemical record of deep-sea sediments provide two indirect methods for investigating the nature of deglaciation and sea-level rise, as opposed to directly measuring the decrease in ice volume in the polar regions. Detailed $\partial^{18}O$ stratigraphy from deep sea cores suggest the late Wisconsinan-Holocene deglaciation proceeded in steps or "terminations" (as many as three have been proposed; Ruddiman and Duplessy, 1985) which may reflect periods of rapid introduction of ^{16}O enriched water to the oceans and/or cooling of the oceans (Chappell and Shackleton, 1986). The use of the $\partial^{18}O$ signature as a proxy for ice volume is not accepted unreservedly because the temperature effect alone could account for the terminations observed (Ruddiman, 1987). These terminations represent changes of 0.5 $\partial^{18}O$ and

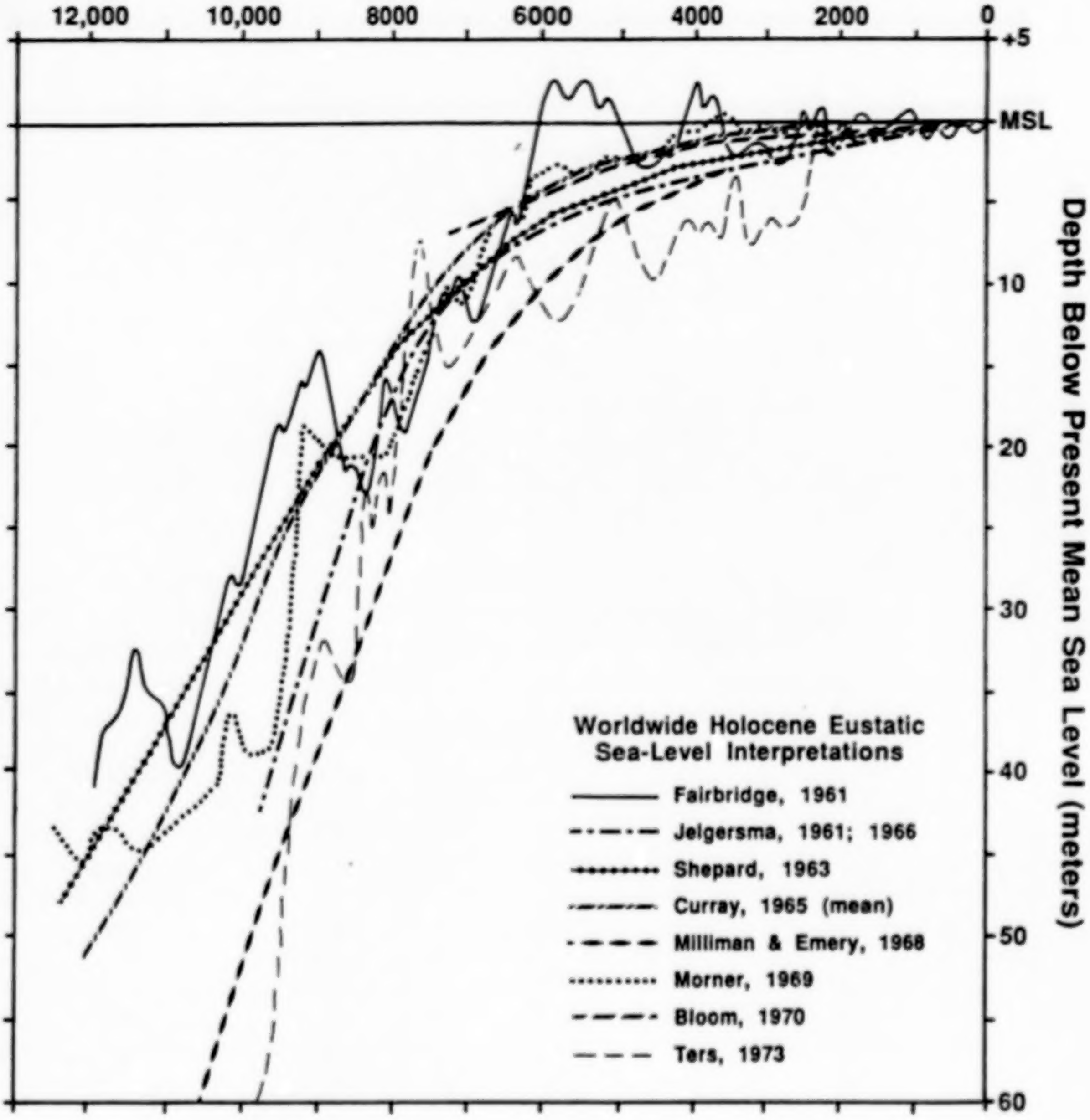


Figure 8. Compilation of Holocene sea level curves from various global locations. Curve shapes reflect the two major schools of thought regarding the nature of the latest transgression: smooth and continuous versus stepped and discontinuous (from Belknap and Kraft, 1977).

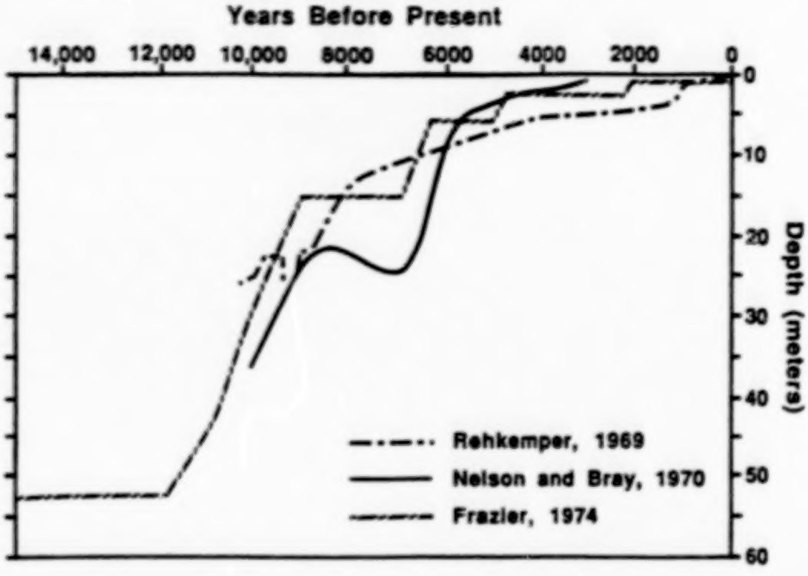


Figure 9. Holocene sea level curves for the Gulf of Mexico region, which interpreted the transgression as proceeding in a series of rises and stillstands.

could translate to a sea-level rise on the order of 35 m (the total glacial to interglacial response is 1.5 $\partial^{18}O$ for ≈ 100 m of sea-level rise). In comparison, the theoretical sea level response of a single marine ice sheet decoupling event is on the order of 5 to 10 m, which corresponds to .05 to .10 $\partial^{18}O$, a magnitude of change that is near or below instrument resolution. Therefore, the $\partial^{18}O$ record is not at this time a practical method for either documenting or disproving the occurrence of high-frequency, low-amplitude sea-level changes. Given this, the stratigraphic record on continental shelves provides the best hope for obtaining the high resolution sea level record needed to address the questions raised by SeaRISE.

Conventionally, sea level curves have been inferred from the age and altitude relationship of sequential shoreline features. A problem with this approach is that transgressive stratigraphy is not due solely to rising sea-level, but can also result from decreased sedimentation rate or increased subsidence rate. To constrain the variables of sedimentation and subsidence rates one should compare the sea-level records from a variety of coastal settings.

Another method for obtaining sea level-curves involves analysis of faunal and floral assemblage changes in carbonate sequences for evidence of changing water depth. This method has recently been used in coastal exposures of Barbados with good success (Fairbanks, 1990). If this method is to be applied to address the questions of SeaRISE, water depth resolution must be improved to a few meters.

Research describing the effects of hydro-isostasy and geoidal influences has led to the consideration of late Wisconsinan-Holocene sea-level curves as regional phenomena (Peltier, 1980; Kidson, 1982), thereby discounting the use of regional sea-level curves as a eustatic indicator. However, the duration of an ice-sheet decoupling/sea-level rise event may be only several hundred years (Clark and Lingle, 1979; Thomas and Bentley, 1978), whereas the viscous response time of hydro-isostasy is several thousand years (Peltier, 1980). Furthermore, Carter et al. (1986) suggest that in middle to low latitude regions correlation by altitude alone is possible, because hydro- and glacio-isostatic effects will be negligible there. This is especially true during the period from 18,000 to 6,000 yrs. B.P.; for according to viscous mantle response models (Peltier, 1980; Clark et al., 1986), most of the differential sea-level behavior occurs only after deglaciation is complete. It is probable then, that high-frequency, low-amplitude sea-level rises due to marine ice-sheet decoupling may be superimposed on the gradual changes in sea level associated with climatic warming and isostatic adjustment. The question is, how do we obtain sea-level curves that are accurate enough to show whether such changes have occurred?

Thomas (1990) and Anderson and Thomas (in press) contend that there is a pattern of episodic sea-level rise worldwide that transcends the six zones of sea-level response (due to viscous mantle behavior) as defined by Peltier (1980). They cite examples of sea-level curves from around the globe that provide evidence for rapid, episodic rises and argue that a pattern for global correlation of such events is emerging. For example, Tooley (1978) found evidence for a 5 m rise that occurred between 7,800 and 7,600 yr B.P., which he attributed to collapse of the Laurentide ice sheet. Carter et al. (1986) documented periods of rapidly rising sea level throughout the last transgression based on studies conducted in Australia and New Zealand. Also, an episodic relative sea-level rise constitutes the favored model for the late Wisconsinan-Holocene transgression in the Gulf of Mexico region (Frazier, 1974; Penland et al., 1990). The problem is that the methods used to derive these curves generally are not capable of resolving low amplitude changes of the scale caused by rapid mass wasting of marine ice sheets. Furthermore, most of the curves shown in Figure 8 rely on radiometric ages of either shell material dredged from banks or peats collected in cores. The assumption that these are shoreline deposits is, more often than not, poorly supported. Thomas (1990) demonstrates that these curves are in error in the case of the Gulf of Mexico (Fig. 9) .

A key obstacle to acquiring high resolution sea-level curves from clastic shelves stems from the fact that Holocene deposits on continental shelves are characteristically thin (generally less than 1 m) and almost always reworked by marine organisms and/or marine processes (Curray, 1960). Thick, undisturbed Holocene strata are required to develop reliable sea-level curves. Belknap and Kraft (1981) contend that a very good record of the late Wisconsinan-Holocene transgression is preserved within incised valley-fill deposits, which escape the effects of shoreface erosion. Thus, these sequences are thick (generally tens of meters) and complete. Because they are thick, these sequences provide the only opportunity for obtaining the high resolution seismic records needed to document shoreline changes.

Ongoing research on the north Texas shelf is aimed specifically at examining the stratigraphic record for rapid, episodic sea-level events (Thomas and Anderson, 1988, Thomas, 1990; Anderson and Thomas, in press). This study focuses on the Trinity/Sabine valley system. High resolution seismic data from the incised valley system, integrated with lithologic data from piston cores, vibracores, and oil industry geotechnical boring descriptions, allow the mapping of estuarine and coastal systems that occupy the valley (Fig. 10). Sea-level stillstands are inferred from paired upper-bay marsh and tidal-inlet estuarine systems (Fig. 11). So far, three such systems, at -14 m, -20 m, and -29 m, have been defined seismically. Large sand bodies, associated with the tidal complexes, most likely were coeval shoreline deposits; however, the shoreline sands (used to recreate sea-level history in the Gulf of Mexico; Curray, 1960; Nelson and Bray, 1970; and Frazier, 1974) were reworked and moved landward. Rapid rises in sea level are manifested in the form of back-stepping parasequences (Fig. 11) which reflect times when fluvial/estuarine/marine systems transgressed tens of kilometers up valley along virtually flat surfaces. It was during these rapid rises that coastal barriers were overstepped to create offshore sand bodies.

The next step in this investigation will be to acquire cores from units that will provide radiometric dates for these sea-level events. But before these events can be accepted as global in scale, they must be correlated with records from similar studies in other parts of the world. An added bonus from such studies is that they provide a record of the impact that rapid sea-level changes have had on coasts and estuaries. Perhaps accurate dating of decoupling events in the Arctic or Antarctic will permit direct correlation of cause and effect. SeaRISE may provide such data.

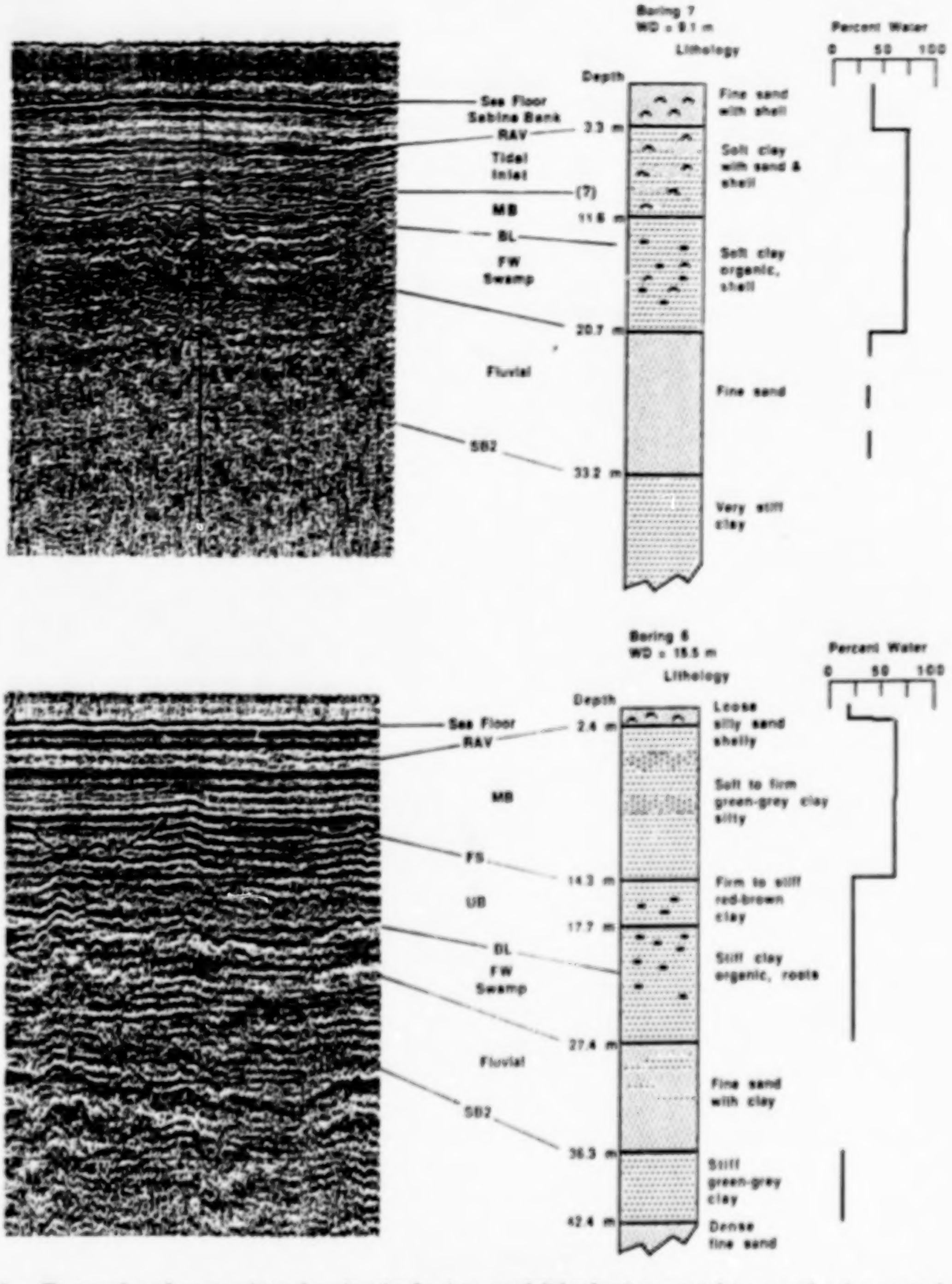


Figure 10. Example of associated seismic facies and lithofacies used to study the sea level record of the Trinity/Sabine incised valley. The interpreted seismic record at the top (a) shows a valley fill sequence formed during the slow rise/still stand of the past 2500 years. Note that fluvial deposits are overlain by estuarine deposits and these ,in turn, are overlain by tidal inlet/flood tidal delta deposits. The interpreted seismic record at the bottom (b) was acquired in the offshore Sabine Valley and within a segment of the valley where rapid flooding took place. This is indicated by the absence of tidal inlet/flood tidal delta facies and by seismic records collected up the valley axis and through the location of this profile that show a back-stepping parasequence boundary between the estuarine and marine deposits. WD = water depth; RAV = ravinement surface; MB = middle bay; FS = flooding surface; UB = upper bay; BL = bay line; FW Swamp = fresh water swamp; SB2 = sequence boundary 2.

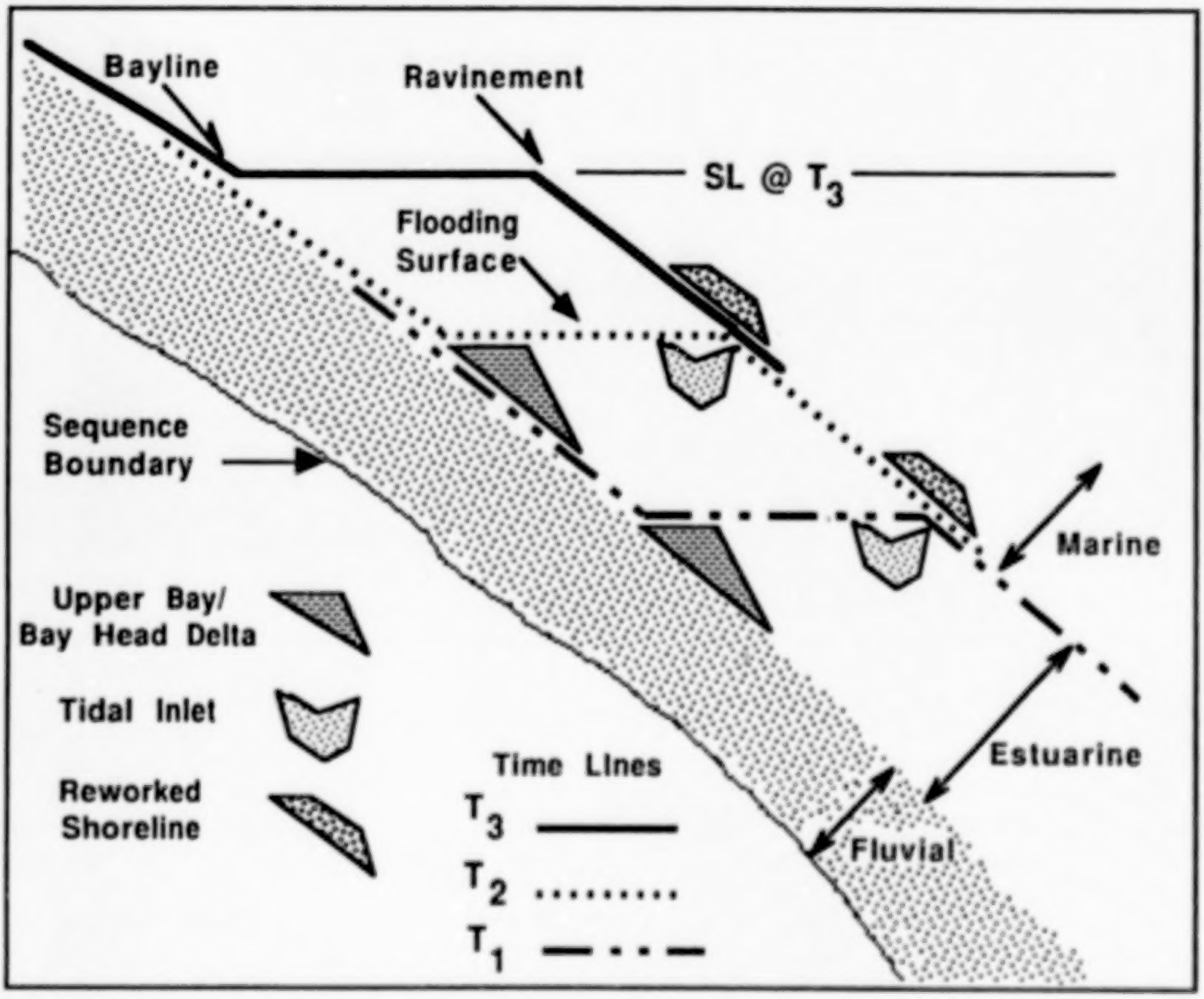


Figure 11. Model depicting the method used to measure sea level changes from high resolution seismic records (from Thomas, 1990). During still stands and slow rises, paired bay-head delta/flood tidal delta-tidal inlet facies develop. During rapid rises, these systems are overstepped and the flooding surface is manifested as a back-stepping parasequence boundary.

REFERENCES CITED

Alley, R.B., Blankenship, D.D., Rooney, S.T., and Bentley, C.R., 1989, Sedimentation beneath ice shelves-The view from ice stream B: Marine Geology, v. 85, p. 101-120.

Anderson, J.B., Domack, E.W., Kennedy, D., and Smith, M., in press, Sedimentary facies associated with Antarctica's floating ice masses: in Anderson, J.B. and Ashley, G.M., eds., Paleoclimatic Interpretation from Glacial Marine Deposits, Geological Society of America Special Publication.

Anderson, J.B, Kurtz, D.D., Domack, E.W., and Balshaw, K.M., 1980, Glacial and glacial marine sediments of the Antarctic continental shelf: Journal of Geology, v. 88, p. 399-414.

Anderson, J.B., Brake C., and Myers N., 1984, Sedimentation on the Ross Sea continental shelf, Antarctica: Marine Geology, v. 57, p. 295-333.

Anderson, J.B. and Thomas, M.A., in press, Marine ice sheet decoupling as a mechanism for rapid, episodic sea level change: the record of such events and their influence on sedimentation: Marine Geology.

Balshaw, K.M., 1981, Antarctic glacial chronology reflected in the Oligocene through Pliocene sedimentary section in the Ross Sea: Unpublished Ph.D. Dissertation, Rice University, Houston, Texas, 140p.

Bartek, L.R., 1989, Sedimentology and stratigraphy of McMurdo Sound and the Ross Sea: implications for Cenozoic glacial history and basin analysis of high latitude marginal basins: Unpublished Ph.D. Dissertation, Rice University, Houston, Texas, 416p.

Bartek, L.R. and Anderson, J.B., 1990, Interpretation of glacial marine deposits and the stratigraphic record of Antarctica's glacial, climatic, and oceanographic history: Problems, progress, and future needs: in Cooper, A.K. and Webb, P.N., eds., International Workshop on Antarctic Offshore Seismic Stratigraphy (ANTOSTRAT): Overview and Extended Abstracts, U.S. Geological Survey Open-File Report 90-309, p. 25-30.

Bartek, L.R. and Anderson, J.B., in press, Facies distribution resulting from sedimentation under polar interglacial climatic conditions within a high latitude marginal basin, McMurdo Sound, Antarctica: in Anderson, J.B. and Ashley, G.M., eds., Paleoclimatic Interpretation from Glacial Marine Deposits, Geological Society of America Special Publication.

Belknap, D.F. and Kraft, J.C., 1977, Holocene relative sea-level changes and coastal stratigraphic units on the northwest flank of the Baltimore Canyon Trough geosyncline: Journal of Sedimentary Petrology, v. 47, p. 610-629.

Belknap, D.F. and Kraft, J.C., 1981, Preservation potential of transgressive coastal lithosomes on the U.S. Atlantic shelf: in Nittrouer, C.A., ed., Sedimentary Dynamics of Continental Shelves, Elsevier, Amsterdam, p. 429-442.

Bloom, A.L., 1970, Paludal stratigraphy of Truk Ponape and Kusaie, eastern Caroline Islands: Geological Society of America Bulletin, v. 81, p. 1895-1904.

Carter, R.M., Carter, L., and Johnson, D.P., 1986, Submerged shorelines in the SW Pacific: evidence for an episodic post-glacial transgression: Sedimentology, v. 33, p. 629-649.

Chappell, J. and Shackleton, N.J., 1986, Oxygen isotopes and sea level: Nature, v. 324, p. 137-140.

Chriss, T. and Frakes, L.A., 1972, Glacial marine sedimentation in the Ross Sea: in Adie, R.J., ed., Antarctic Geology and Geophysics, Commission on Antarctic Research, Oslo, p. 747-762.

Clark, J.A., and Lingle, C.S., 1979, Predicted sea-level changes (18,000 years b.p. to present) cuased by late glacial retreat of the Antarctic ice sheet: Quaternary Research, v. 11, p. 279-298.

Clark, J.A., Farrel, W.E., and Peltier, W.R., 1978, Global changes in postglacial sea level: a numerical calculation: Quaternary Research, v. 9, p. 265-287.

Curray, J.P., 1960, Sediments and history of the Holocene transgression, continental shelf, Northwest Gulf of Mexico: in Shepard, F.P., Phleger, F.B., and Van Andel, T.H., eds., Recent Sediments, Northwest Gulf of Mexico: American Association of Petroleum Geologists, Tulsa, Oklahoma, p. 221-266.

Curray, J. R., 1965, Late Quaternary history, continental shelves of the United States: in Wright, H.E. and Frey, D.E., eds., The Quaternary of the United States, Princeton University Press, Princeton, p. 723-735.

David, T.W.E. and Priestley, G.E., 1914, Glaciology, physiography, stratigraphy, and tectonic geology of South Victoria Land: British Antarctic Expedition, 1907-1909: Reports of the Scientific Investigations, Geology, 1.

Debenham, F., 1921, Recent and local deposits of McMurdo Sound region: Natural History Report, British Museum, British Antarctic Expedition, Geology, v. 1, p. 63-90.

Denton, G.H. and Hughes, T.J., 1981, The Last Great Ice Sheets, John Wiley and Sons, New York, 484p.

Domack, E.W., 1982, Sedimentology of glacial and glacial-marine deposits on the George V-Adelie continental shelf, East Antarctica: Boreas, v. 11, p. 79-97.

Domack, E.W., Anderson, J.B., and Kurtz, D.D., 1980, Clast shape as an indicator of transport and depositional mechanisms in glacial marine sediments: George V Continental Shelf, Antarctica: Journal of Sedimentary Petrology, v. 50, p. 813-820.

Domack, E.W. and Jull, A.J.T., in press, Holocene chronology for the unconsolidated sediments at Site 740A: Prydz Bay, East Antarctica: in Barron, J. and Larsen, B., eds., Proceedings of the Ocean Drilling Program, Scientific Results, College Station, Texas, v. 119B.

Domack, E.W., Jull, A.J.T., Anderson, J.B., Linick, T.W., and Williams, C.R., 1989, Application of tandem accelerator mass-spectrometer dating to late Pleistocene-Holocene sediments of the East Antarctic continental shelf: Quaternary Research, v. 31, p. 277-287.

Drewry, D.J., 1979, Late Wisconsin reconstruction for the Ross Sea region, Antarctica: Journal of Glaciology, v. 24, p. 231-244.

Drewry, D.J., 1983, Antarctica: Glaciological and Geophysical Folio, Scott Polar Research Institute, Cambridge.

Dunbar, R.B., Anderson, J.B., Domack, E.W., and Jacobs, S.S., 1985, Oceanographic influences on sedimentation along the Antarctic continental shelf: in Jacobs, S.S., ed., Oceanology of the Antarctic Continental Shelf, American Geophysical Union, Antarctic Research Series, v. 43, p. 291-312.

Edwards, B.D., Lee, H.J., Karl, H.A., Reimnitz, E., and Timothy, L.A., 1987, Geology and physical properties of Ross Sea, Antarctica, continental shelf sediment: in Cooper, A.K. and Davey, F.J., eds., The Antarctic Continental Margin: Geology and Geophysics of the Western Ross Sea, Circum-Pacific Council for Energy and Mineral Resources Earth Science Series, v. 5B, p. 191-216.

Fairbanks, R.G., 1990, A 17,000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation: Nature, v. 342, p. 637-642.

Fairbridge, R.W., 1961, Eustatic changes in sea-level: Physics and Chemistry of the Earth, v. 5, p. 99-185.

Fillon, R. J., 1975, Late Cenozoic paleo-oceanography of the Ross Sea, Antarctica: Bulletin of the Geological Society of America, v. 86, p. 839-845.

Frazier, D.E., 1974, Depositional Episodes: their relationship to the Quaternary stratigraphic framework in the Northwest portion of the Gulf Basin: Texas Bureau of Economic Geology, Geological Circular 74-1, 28p.

King, L.H. and Fader, G.B.J., 1986, Wisconsinan glaciation of the Atlantic continental shelf of southeast Canada, Geological Society of Canada, Geological Survey Bulletin, 363, 72p.

Frazier, D.E., 1974, Depositional Episodes: their relationship to the Quaternary stratigraphic framework in the Northwest portion of the Gulf Basin: Texas Bureau of Economic Geology, Geological Circular 74-1, 28p.

Harwood, D.M., Scherer, R.P., and Webb, P-N., 1989, Multiple Miocene marine productivity events in West Antarctica as recorded in Upper Miocene sediments beneath the Ross Ice Shelf (Site J-9): Marine Micropaleontology, v. 15, p. 91-115.

Hayes, D.E. and Frakes, L.A., 1975, General Synthesis: Deep Sea Drilling Project 28: in Hayes, D.E. and Frakes, L.A., eds., Initial Reports of the Deep Sea Drilling Project, U.S. Government Printing Office, Washington, D.C., v. 28, p. 919-942.

Houtz, R.E. and Davey, F.J., 1973, Seismic profiler and sonobuoy measurements in the Ross Sea, Antarctica: Journal Geophysical Research, v. 78, p. 3448-3468.

Houtz, R.E. and Meijer, R., 1970, Structure of the Ross Sea shelf from profiler data: Journal of Geophysical Research, v. 75, p. 6592-6597.

Hughes, T., 1973, Is the West Antarctic Ice Sheet disintegrating?: Journal of Geophysical Research, v. 78, p. 7884-7910.

Hughes, T., 1977, West Antarctic ice Streams: Review of Geophysics and Space Physics, v. 15, p. 1-46.

Hughes, T.J., 1987, Deluge II and the continent of doom: rising sea level and collapsing Antarctic ice: Boreas, v. 16, p. 89-99.

Imbrie, J., Hays, J.D., Martinson, D., McIntyre, A., Mix, A., Morley, J., Pislas, N., Prell, W., and Shackleton, N.J., 1984, The orbital theory of Pleistocene climate: support from a revised chronology of the marine $d^{18}O$ record: in Berger, A.L. et al., eds., Milankovitch and Climate, Part I, p. 269-305.

Jacobs, S.S., Fairbanks, R.G., and Horibe, Y., 1985, Origin and evolution of water masses near the Antarctic continental margin: evidence from $H_2{}^{18}O/H_2{}^{16}O$ ratios in seawater: in Jacobs, S.S., ed., Oceanology of the Antarctic Continental Shelf, American Geophysical Union, Antarctic Research Series, v. 43, Washington, D.C., p. 59-86.

Jelgersma, S., 1961, Holocene sea-level changes in the Netherlands: Med. Geol. Stich., C-VI, 101 p.

Jelgersma, S., 1966, Sea-level changes during the last 10,000 years: Proceedings of the International Symposium on World Climates from 8000 B.C. to 0 B.C. Imperial College, London, Royal Meterological Society, p. 54-71.

Karl, H.A., Reimnitz, E., and Edwards, B.D., 1987, Extent and nature of Ross Sea Unconformity in the western Ross Sea, Antarctica: in Cooper, A.K. and Davey, F.J., eds., The Antarctic Continental Margin: Geology and Geophysics of the Western Ross Sea, Circum-Pacific Council for Energy and Mineral Resources Earth Science Series, 5B, p. 77-92.

Kellogg, D.E. and Kellogg, T.B., 1986, Diatom biostratigraphy of sediment cores from beneath the Ross Ice Shelf: Micropaleontology, v. 32, p. 74-94.

Kellogg, T.B. and Truesdale, R.S., 1979, Late Quaternary paleoecology and paleoclimatology of the Ross Sea: The diatom record: Marine Micropaleontology, v. 4, p. 137-158

Kellogg, T.B., Truesdale, R.S., and Osterman, L.E., 1979, Late Quaternary extent of the West Antarctic Ice Sheet: New evidence from Ross Sea cores: Geology, v. 7, 249-253.

Kendall, C.G.St. C. and Lerche, I., 1988, The rise and fall of eustasy: in Wilgus, C.K., Hastings, B.S., Kendell, C.G. St. C., Posamentier, H.W., Ross, C.A., and Van Wagoner J.C., eds., Sea-Level Changes: An Integrated Approach, Society of Econonic Paleontologists and Mineralogists Special Publication No. 42, Tulsa, p. 3-18.

Kennedy, D.S and Anderson, J.B., 1989, Quaternary glacial history of Marguerite Bay, Antarctic Peninsula: Quaternary Research, v. 31, p. 1-22.

Kidson, C., 1982, Sea level changes in the Holocene: Quaternary Science Reviews, v. 1, p. 121-151.

Kurtz, D.D. and Anderson, J.B., 1979, Recognition and sedimentologic description of recent debris flow deposits from the Ross and Weddell Seas, Antarctica: Journal of Sedimentary Petrology, v. 94, p. 1157-1170.

Ledford-Hoffman, P.A., DeMaster, D.J., and Nittrouer, C.A., 1986, Biogenic-silica accumulation in the Ross Sea and the importance of Antarctic continental-shelf deposits in the marine silica budget: Geochemica et Cosmochimica Acta, v. 50, p. 2099-2110.

Leventer, A. and Dunbar, R.B., 1988, Recent diatom record of McMurdo Sound, Antarctica: Implications for history of sea ice extent: Paleoceanography, v. 3, p. 259-274.

Mayewski, P.A., 1975, Glacial geology and late Cenozoic history of the Transantarctic Mountains, Antarctica: Institute of Polar Studies Report No. 56, 168p.

Milliman, J. D. and Emery, K. O., 1968, Sea levels during the past 35,000 years: Science, v. 162, p. 1121-1123.

Morner, N. A., 1969, The Late Quaternary history of the Kattegatt Sea and the Swedish west coast: Sveriges Geol. Under., C640, 487p.

Myers, N.C., 1982, Petrology of Ross Sea basal tills: Antarctic Journal of the United States, v. 17, p. 123-124.

Myers, N.A., 1982, Marine geology of the western Ross Sea: implications for Antarctic glacial history, Unpublished MA thesis, Rice University, Houston, Texas, 234 p.

Nelson, H.F. and Bray, E.E., 1970, Stratigraphy and history of the Holocene sediments in the Sabine-High Island area, Gulf of Mexico: in Morgan, J.P., ed., Deltaic Sedimentation Modern and Ancient: Society of Econonic Paleontologists and Mineralogists Special Publication No. 15, Tulsa, p. 48-77.

Peltier, W.R., 1980, Models of glacial isostacy and relative sea-level: in Bally et al., A.W., eds., Dynamics of Plate Interiors, Geodynamics Series, v. 1, American Geophysical Union, Washington, D.C., p. 111-128.

Penland, S., Suter, J.R., and McBride, R.A., 1988, Delta-plain development and sea level history in the Terrebonne coastal region, Louisiana: in S. Penland, S. and Suter, J.S., eds., Coastal and Shallow Marine Sedimentation on the Mississippi River Delta Plain and Chenier Plain: Society of Econonic Paleontologists and Mineralogists Field Trip No. 2 Guidebook, p. 2-23.

Penland, S, Suter, J.R., Boyd, R., and Williams, S.J., 1990, Effects of sea level rise on river delta coasts (abstract): American Association of Petroleum Geologists Bulletin, v. 74, p. 738.

Priestley, R.E., 1923, Physiography (Robertson Bay and Terra Nova Bay regions): British Antarctic (Terra Nova) Expedition, 1910-1913, Harrison and Sons, London.

Rehkemper, L.J., 1969, Sedimentology of Holocene estuarine deposits, Galveston Bay: in Lankford, R.R. and Rodgers, J.J.W., eds., Holocene Geology of the Galveston Bay Area, Houston Geological Society, p. 12 - 52.

Reid, D. E., 1989, Late Cenozoic glacial-marine, carbonate, and turbidite sedimentation in the northwestern Ross sea, Antarctica: Unpublished Master's Thesis, Rice University, Houston, Texas, 178p.

Ruddiman, W.F., 1987, Synthesis; the ocean/ice sheet record: in Ruddiman, W.F. and Wright, H.E., Jr., eds., North America and Adjacent Oceans During the Last Deglaciation, Decade of North American Geology, v. K-3, Geological Society of America, Boulder, p. 463 - 478.

Ruddiman, W.F. and Duplessy, J.-C., 1985, Conference on the last deglaciation; Timing and mechanism: Quaternary Research, v. 23, p. 1-17.

Savage, M.L. and Ciesielski, P.F., 1983, A revised history of glacial sedimentation in the Ross Sea region, in Oliver, R.L., James, P.R., and Jago, J.B., eds., Antarctic Earth Science, Australian Academy of Science, Canberra, p. 555-559.

Scott, R.F.,1905, The Voyage of the Discovery, Charles Scribner's Sons, 2 volumes, 556p and 508p.

Shennan, I., 1987, Global analysis and correlation of sea-level data: in Devoy, R.J.N., ed., Sea Surface Studies, a Global View, Croom Helm, New York, p. 197-232.

Shepard, F.P., 1931, Glacial troughs of the continental shelves: Journal of Geology, v. 39, p. 345-360.

Shepard, F. P., 1963, 35,000 years of sea level: in Clements, T., ed., Essays in Marine Geology, University of Southern California Press, Los Angeles, p. 1-10.

Stuiver, M., Denton, G.H., Hughes, T.J., and Fastook, J.L., 1981, History of the marine ice sheet in West Antarctica during the last glaciation: a working hypothesis: in Denton, G.H. and Hughes, T.J., eds., The Last Great Ice Sheets, Wiley, New York, p. 319-439.

Ters, M., 1973, Les variations du niveau marin depuis 10, 000 ans le long du littoral Atlantique Francais: in Le Quaternaire: geodynamique, stratigraphie et environment, travaux Francais recents, 9th Congres International de l'I.N.Q.U.A., Christchurch, New Zealand, p. 114-135.

Thomas, M.A., 1990, The impact of long-term and short-term sea level changes on the evolution of the Wisconsinan-Holocene Trinity-Sabine incised valley system, Texas continental shelf: Unpublished Ph.D. Dissertation, Rice University, Houston, Texas, 295p.

Thomas, M.A. and Anderson, J.B., 1988, The effect and mechanism of episodic sea level events: the record preserved in late Wisconsinan-Holocene incised valley-fill sequences: Transactions of the Gulf Coast Geological Societies, v. 38, p. 399-406.

Thomas, R. H. and Bentley, C.R., 1978, A model for Holocene retreat of the West Antarctic Ice Sheet: Quaternary Research, v. 10, p. 150-170.

Tooley, M.J., 1978, Sea level changes, North-West England during the Flandrian Stage, Clarendon Press, Oxford, 232p.

Truswell, E.M. and Drewry, D.J., 1984, Distribution and provenance of recycled palynomorphs in surficial sediments of the Ross Sea, Antarctica: Marine Geology, v. 59, p. 187-214.

Vanney, J.R., Falconer, R.K.H., and Johnson, G.L., 1981, Geomorphology of the Ross Sea and adjacent oceanic provinces: Marine Geology, v. 41, p. 73-102.

Whillans, I.M., 1976, Radio-echo layers and the recent stability of the West Antarctic ice sheet: Nature, v. 264, p. 152-155.

Wong, H.K. and Christoffel, D.A., 1981, A reconnaissance seismic survey of McMurdo sound and Terra Nova Bay, Ross Sea: in McGinnis, L.D., ed., Dry Valley Drilling Project, Antarctic Research Series 33, Washington, D.C., p. 37-62.

Ice Cores and SeaRISE--What We Do (and Don't) Know

Richard B. Alley

Earth System Science Center and Department of Geosciences
The Pennsylvania State University
University Park, PA 16802, U.S.A.

ABSTRACT. Ice-core analyses are needed in SeaRISE to learn what the West Antarctic ice sheet and other marine ice sheets were like in the past, what climate changes led to their present states, and how they behave. The major results of interest to SeaRISE from previous ice-core analyses in West Antarctica are that the end of the last ice age caused temperature and accumulation-rate increases in inland regions, leading to ice-sheet thickening followed by thinning to the present.

INTRODUCTION.

The goal of SeaRISE is to make predictions of marine-ice-sheet changes and any contribution they may make to changes in global sea level, based on scenarios of the future produced by other global-change research. For example, a scenario might be "If humans burn all of the known oil and coal reserves on the planet over the next 100 years and raise the global temperature 5°C", and the prediction might be "then the West Antarctic ice sheet (WAIS) will collapse, raising global sea level 6 m in 200 a". Different scenarios might lead to different predictions.

The predictions must be made with models--simplified representations of the real marine-ice-sheet system. All models, whether numerical, analytic, or analog, share certain features. These features include a description or representation of how the system being modeled behaves (governing equations in a numerical model), what the system is like at some time (initial conditions), and what is done to it by things outside (boundary conditions).

In addition, for a model to be useful, it must be tested or validated. For earth processes that occur over decades or centuries, the only practical way to test a model is against data from the past: if the model can start from some former conditions and, using past forcings, predict the present state (or if the model can "retrodict" past states), then we can place some confidence in its ability to predict the future.

In building the SeaRISE model(s), ice cores are critical for learning ice-sheet history for validation, and play an important role in learning the governing equations and even the initial conditions. Isotopic, trace-element and total-gas analyses of ice cores contribute primarily to the history, and physical-property measurements provide information on the governing equations. In addition, the borehole left by core recovery is valuable for extracting history and initial conditions from ice temperatures, learning governing equations from borehole deformation and logging, and gaining access to the bed for further studies.

Here, I briefly review what is known, and not known, about marine ice sheets (primarily the WAIS) based on ice-core measurements. I provide some literature citations to guide the interested reader, but make no attempt to present a fully referenced review. In order, I discuss ice-core contributions to learning ice-sheet history (global record, surface temperature, accumulation rate, surface elevation, and other information), current conditions, and governing equations, and I conclude with an editorial on what further studies are needed for SeaRISE.

ICE-SHEET HISTORY.

THE GLOBAL RECORD.

An ice core records both local and global events. Of these, local events are more valuable to SeaRISE, but the global signal may be of greater interest to the broader scientific community.

Probably the most notable result from ice-core research has been the demonstration that the ice-age atmosphere contained less CO_2 and CH_4, and more dust, than the modern atmosphere (Thompson and others, 1981; Lorius and others, 1989; Chappellaz and others, 1990). Reduction of these greenhouse gases, and increase of dust, would have tended to cool the planet, contributing to the low ice-age temperatures (Harvey, 1988). The changes may have been startlingly rapid, although ice flow has caused at least some of the changes to appear more rapid than they actually were (Staffelbach and others, 1988).

These and similar results place significant constraints on global climate models, and provide important clues to the past operation of the global climate system. To the extent that marine ice sheets are embedded in the world climate, these results are useful in determining ice-sheet history. However, they are unlikely to be of direct, primary importance to SeaRISE.

TEMPERATURE.

Of greater interest to SeaRISE are ice-core analyses for more local effects: near-surface temperature, net snow accumulation, and ice-surface elevation. These require measurement of isotopes, total gas content, physical properties, and annually varying properties in ice cores, combined with borehole measurements, atmospheric and oceanic modeling, and ice-flow modeling.

For temperatures, the most useful information for SeaRISE is the history of the near-surface snow temperature (the 10-m temperature), not the air temperature. The rate of ice deformation is known to increase exponentially with the ice temperature, which changes over climatic cycles. The rate at which changes in 10-m temperature propagate into the ice and affect its flow is readily calculated if the flow pattern is known, so the 10-m temperature provides a useful boundary condition for ice-flow modeling. The 10-m temperature may differ from the mean annual near-surface air temperature by a few tenths of a degree to a few degrees, however, depending on radiative conditions, wind pumping, rates of sublimation, and other factors that are difficult to model or measure, and that may have changed over time (Paterson, 1981, ch. 10). (Much is known about these processes, but we cannot calculate the difference between air and snow temperatures from first principles, nor can we estimate how it has changed with time.)

The only direct sensor of past 10-m temperatures is the physical temperature of the ice today, and of underlying bedrock if the ice is frozen to its bed. Any change in snow temperature moves downward into the ice sheet by conduction and ice flow. If the ice flow, heat sources, and thermal properties of ice are known, then the steady-state temperature profile in ice can be calculated. Deviations of measured temperatures from the steady-state profile represent climatic changes, with the age of the changes increasing with the depth of the deviations in the ice.

Because of thermal diffusion, the deviation of measured temperature from steady state caused by a surface-temperature change decreases in amplitude but spreads over a greater depth range with increasing age, and eventually becomes too small to measure or overlaps perturbations of similar age and ceases to be recognizable. Thus, whereas seasonal cycles of the last year or two are evident in temperature records from the upper few meters, events thousands of years old had to persist for hundreds of years to be recognizable in temperature records today.

Measurement of ice temperatures to 0.01°C or better is possible once a hole exists and has equilibrated with the surrounding ice, and such accuracy is needed for paleoclimatic interpretation. Such accuracies are difficult to achieve from thermistor strings frozen into holes melted with hot water, owing to thermistor drift and freeze-in pressure. Hot-water drilling combined with antifreeze and casing could be developed to allow temperature measurements without core drilling (G. Clow, private communication, 1990), but the only proven method today for accurate, deep measurements is to utilize a hole left by ice-core drilling.

Most earlier workers have tested whether the ice-temperature record is consistent with some assumed history of near-surface snow temperatures, by using that assumed history to drive a forward model of ice temperature and asking whether this explained most of the variance between observed and steady-state model profiles (within unspecified but rather broad error limits; e.g. Johnsen, 1977). The assumed history may be derived from isotope records or a variety of regional proxy records and may be adjusted in a more-or-less *ad hoc* method to optimize the fit between model and observed profiles (e.g. Fig. A; Alley and Koci, 1990). A better approach is to use inverse theory, a technique that allows determination of that surface-temperature history (or histories?) most consistent with the observed temperature profile, and that is independent of other proxy records (MacAyeal and others, in review).

The major results of borehole temperature analyses to date have been to demonstrate the basic correctness of isotopic or other proxy records (e.g. Dahl-Jensen and Johnsen, 1986), and to provide information on ice-flow patterns (Bolzan, 1985). It certainly should be possible to obtain better results from new temperature records and from re-analysis of previously measured temperatures, using the more-sophisticated analysis tools now under development. The major drawbacks in use of modern ice temperatures to infer past 10-m temperatures are: 1) ice dynamics and past accumulation rates must be known to calculate temperature profiles for comparison with measured data; 2) the method lacks high resolution for old events; and 3) analyses to date have not been optimized.

An independent method to infer past temperatures, and the method that has been used most widely, is to calculate the temperatures from measurements of oxygen- or hydrogen-isotope ratios in ice cores.

The isotopic composition of snowfall depends on: the composition of the source waters; the degree of equilibrium and nonequilibrium isotopic fractionation during evaporation from the source; the amount of water-vapor removal and the temperature history of that removal prior to formation of the precipitation of interest; and the condensation temperature of the precipitation of interest. If the source and atmospheric-path terms are known, then the isotopic composition gives the condensation temperature for the snowfall. Assuming further that the condensation temperature changes in the same way over time as the near-surface air temperature, the isotopic composition is a proxy for the air temperature (Robin and Johnsen, 1983). Finally, if snow drifting is limited, if sublimation following deposition does not change the isotopic composition (except perhaps by diffusional smoothing of annual signals), if the seasonality of precipitation does not change over time, and if the source of the ice in a core can be determined from ice-flow calculations (Fig. B), then the isotopic composition of the ice in a core provides the near-surface air temperature when and where that ice was deposited, which can be taken as a proxy for the 10-m temperature (with the further uncertainties about offset between snow and air temperatures, as discussed above).

The major advantages of this method of paleothermometry over ice-temperature measurements are that the isotopes have much higher resolution (isotopic diffusion does smooth the record with time, but this smoothing is some orders of magnitude slower than the smoothing of the physical temperature record), and that the isotope records from central regions of ice sheets are nearly independent of ice flow, unlike physical temperature records. The major disadvantage is the indirect nature of the measurement: source compositions, evaporation fractionations, condensation histories, offsets between condensation and near-surface air temperatures, diagenetic changes, and offsets between near-surface air temperatures and 10-m temperatures all must be known to provide the glaciologically significant boundary condition.

The situation is not quite so bleak as this may sound. Temperature histories inferred from isotopes are broadly consistent with those from ice temperatures. Isotopic analyses of marine fossils constrain changes of the isotopic composition of oceanic source regions (although with some uncertainty introduced by oceanic temperature changes; Shackelton, 1983). Combined analysis of hydrogen and oxygen isotopes in ice cores provides information on isotopic fractionation during evaporation from the oceans (Jouzel and others, 1982). Atmospheric modeling, including modeling with general circulation models (GCMs), helps estimate isotopic changes prior to the snowfall of interest (Jouzel and others, 1987).

Modern studies show a good correlation between mean annual isotopic composition of snowfall and mean annual near-surface air temperature at a site (Lorius, 1983). This correlation generally is used to convert isotopic compositions to paleotemperatures, thus avoiding the need to calculate path effects from first principles. However, one study comparing isotopic compositions to instrumental records and site temperatures in the Antarctic Peninsula found that the correlation of site temperature with isotopic composition at the present time has a significantly different slope than the correlation between temperature and isotopic composition over time at a site (Peel and others, 1988). Also, if a given isotopic composition is found in Marie Byrd Land in West Antarctica, it indicates a temperature about 10°C warmer than if found in most of East Antarctica (Fig. C, D; Lorius, 1983). The reasons for this difference are not understood fully, so we cannot tell whether it has been maintained over time. These difficulties may not affect interpretation of East Antarctic records, but are a concern in West Antarctica.

If we wish to obtain paleotemperatures from isotopes with accuracy at the 1 or 2°C level, then there is a clear need for improved coupled ocean-atmosphere-sea ice modeling with explicit treatment of oxygen and hydrogen isotopes to learn how climatic factors have affected the isotope-air temperature relation. Further surveys of near-surface isotopic compositions in West Antarctica would be useful. Detailed studies of the processes causing isotopic fractionation are needed. Also, the relation between near-surface air temperature and 10-m snow temperature must be investigated systematically.

Despite the numerous weaknesses, isotopes are generally accepted as the primary paleothermometer for ice sheets, and have provided much of our present knowledge of past temperatures. They indicate that, during the height of the last ice age, surface temperatures in central regions of East and West Antarctica were 5 to 10°C colder than today (Fig. E; Jouzel and others, 1989), with perhaps 10 to 15°C cooling in Greenland (Dansgaard and Oeschger, 1989), and that warming to near modern values occurred over roughly 5000 years in the range of 15,000 to 10,000 years ago. Since this deglacial warming, temperature oscillations have been in the 1 to 2°C range if averaged over more than a few years. Very good isotopic time series are available from a few places (e.g. Mosley-Thompson and others, 1990; Grootes and others, 1990), although the regional coherence and even the climatologic or meteorologic significance of the observed postglacial oscillations are questionable (Benoist and others, 1982; Grootes and others, 1990).

ACCUMULATION RATE.

The second major forcing on ice sheets is net accumulation (or ablation). For the grounded part of the West Antarctic ice sheet, and for most of the East Antarctic and Greenland ice sheets, this is dominated by the surface snowfall. Analysis of ice cores is the only reasonably direct way we have to learn past accumulation, although climate modeling can provide important information and should be pursued.

Net snowfall can be determined from ice cores by measuring annual layer thicknesses and correcting for ice flow, by using flow models to estimate accumulation rates needed to produce observed ice-temperature profiles or surface-elevation histories, or by measuring dilution of some trace constituent with known deposition or production rate. Of these, the methods combining flow modeling with ice-core analyses have been the most practical, but improved resolution from dilution techniques would allow us to learn accumulation independent of ice flow and to use the other techniques to determine ice-flow patterns better.

Many parameters that can be measured in ice cores vary with an annual signal, including various trace chemicals, physical properties, dust, and isotopes (Hammer, 1989). Layer thinning eventually renders all of these methods unusable in ice, and diffusion affects some of them and limits their utility. For isotopes and hydrogen peroxide, an accumulation rate of about 20 cm/a ice or more is needed to allow annual signals to survive diffusion in snow and firn and reach the ice, where diffusion is slower and the signal can be recognized for hundreds or thousands of years (Johnsen and Robin, 1983).

Annual layers have been counted continuously from the surface for several thousand years in Greenland (Dansgaard and Oeschger, 1989), for about 2000 years in East Antarctica (Mosley-Thompson and Thompson, 1982 for 911 years; A.J. Gow, pers. comm., 1990 for 2000 years), and for several hundred years in West Antarctica (Gow, 1968), and older and deeper annual layers have been recognized, although the tedious job of measuring all of the annual layers down to these deeper levels has not been done. Microparticles offer the greatest opportunity for recognizing annual layers in deep and old ice, because they are negligibly affected by diffusion.

The annual-layer thickness provides the accumulation rate only if corrected for density and ice flow. Density is measured easily, and ice flow is relatively well known in the upper half of an ice sheet, especially away from ice divides, so that past accumulation rates can be learned quite accurately for recent times if accumulation rates are high enough to produce annual layers. However, ice flow complicates absolute determination of accumulation rates from deeper ice.

Ice-flow modeling combined with ice-core data can contribute to learning past accumulation rates in other ways. The temperature-depth relation in ice depends on the accumulation rate as well as the surface temperature, and changes in accumulation rate can be estimated by jointly determining the accumulation-rate and surface-temperature histories needed in models to match observed temperature profiles (e.g. Dahl-Jensen and Johnsen, 1986). Similarly, accumulation rate helps control ice-sheet thickness and surface elevation. Past accumulation rates can be estimated by determining the history needed in ice-flow models to match changes in ice-sheet elevation estimated from total-gas-content measurements (see below).

The other direct method of determining accumulation rates from ice-core measurements is to measure dilutions. This technique is in its infancy, but could prove quite useful if developed successfully. The basic idea is that certain trace constituents are deposited on the ice sheet or formed near the surface of the ice sheet at a rate independent of time. If the annual accumulation of snow is large, then the concentration of the trace constituent is low, but if the annual accumulation of snow is small, then the trace constituent is more concentrated. One such constituent is ^{10}Be, which is a cosmogenic isotope with a short atmospheric residence time. With certain exceptions, it is believed that production and deposition rates have been nearly constant over time, allowing changes in snow accumulation to be inferred from concentration changes of ^{10}Be measured in ice cores (Raisbeck and others, 1987).

An indirect method to estimate past accumulation rates is used more often than any of these more-direct methods. At present, net snow accumulation on the East Antarctic ice sheet is inversely correlated with mean annual temperature, in the way that would be expected if accumulation were controlled by the saturation vapor pressure of the air. If it is assumed that saturation vapor pressure of precipitating air masses is correlated with near-surface air temperature and controls mean accumulation rate, then the accumulation rate can be calculated from the surface temperature, which is estimated from isotopes or ice temperatures (Robin, 1977).

Such an approach obviously has certain drawbacks. For example, in the modern world snowfall increases with increasing temperature in central East Antarctica on a mean annual basis, but most snowfall in Antarctica occurs in the winter when it is cold (Bromwich, 1988). In West Antarctica even the mean-annual correlation does not work well in places--Byrd Station is about 1.5°C colder than the Siple Coast, but has almost twice the net accumulation of the Siple Coast (probably owing in part to the anomalously low elevation of the Siple Coast; Robin, 1983; Alley and Bentley, 1988; Whillans and Bindschadler, 1988). Indeed, the variability in accumulation rate for a given temperature in Antarctica approaches an order of magnitude, even if the blue-ice ablation zones are excluded (Robin, 1977). Clearly, physically based atmospheric modeling is needed to test or replace this simple correlation model.

Despite all of the problems of the different methods of inferring past accumulation rates, they do achieve some degree of internal consistency. The temperature-accumulation relation and ^{10}Be produce similar estimates of accumulation during the last ice age in central East Antarctica (Jouzel and others, 1989). In south-central Greenland, model attempts to match the measured physical temperature of the ice require a glacial-interglacial accumulation change similar to that estimated from ^{10}Be, coupled with a temperature change similar to that indicated by isotopes (Dansgaard and Oeschger, 1989). For West Antarctica, changes in surface elevation inferred from measurements of total gas content on the Byrd core indicate a deglacial thickening such as would be produced by a significant increase in accumulation rate (Whillans, 1981; Raynaud and Whillans, 1982), similar to that estimated from layer thicknesses measured by electrical conductivity (Hammer and others, 1985) and estimated from temperature change. In each case, accumulation apparently increased significantly at the end of the last ice age (by about a factor of two, with large error bars). Following this deglacial transition, accumulation seems to have shown relatively small oscillations around a Holocene mean (e.g. Clausen and others, 1988).

Because the most commonly used technique for estimating accumulation rates on glacial-interglacial time scales is to estimate temperature from isotopes, and to estimate accumulation from temperature, we are left with less confidence in our knowledge of accumulation rates than in our knowledge of temperatures. One alternative approach would be to develop good atmospheric models for the ice sheet. If we can determine surface elevations and surface temperatures from ice cores, and use these to constrain atmospheric models, then the models can (in principle) estimate snowfall rates. A second approach would be to accelerate continuing work on the dilution techniques.

A third approach would be to improve the coupling of ice-flow models and ice-core observations. The ice temperature, ice thickness, and form and thickness of layers in the ice depend on the way the ice flows, and on the past surface temperature and accumulation rate (among other factors). A fully coupled model, if used with appropriate inverse techniques, should be able to reveal those combinations of temperature history, accumulation history, and ice-flow parameters that best match the modern state of the ice sheet. The hope is that the result of such a model, constrained by independent data, would yield a unique ice-sheet history. Care must be taken to avoid circularity in such an exercise: we seek paleoclimatic data to test ice-sheet models, but if we use all the data in developing the models, they no longer are testable. Thus, some subset of the data should be held out from model development for testing.

All of these approaches for learning past accumulation rates more accurately merit study. Of them, development of improved models for polar precipitation, and development of coupled ice-flow/ice-core models, seem especially suited for an interdisciplinary program such as SeaRISE.

ICE-SURFACE ELEVATION.

The third major element of ice-sheet history that can be derived from ice-core analysis is surface elevation, and from surface elevation, ice-sheet thickness. Total gas content of the ice is the best measure here, although isotopes also are useful.

In the modern world, the volume of air bubbles per unit volume of ice trapped by the transformation of snow to ice without melting is observed to be a weak function of mean annual temperature and appears to be independent of accumulation rate. From the perfect gas law and this known relation between temperature and bubble volume, the number of air molecules trapped in unit volume of ice yields the air pressure at pore close-off. Because air pressure decreases with increasing elevation, the calculated air pressure can be used to infer past elevations of pore close-off. Further, ice-surface elevations are believed to undergo larger and more-rapid changes than any changes in the depth of pore close-off below the surface, so changes in total gas content are used to estimate changes in surface elevation (Raynaud, 1983).

The method is subject to some uncertainty, of course. The total gas content varies between closely spaced samples, probably indicating seasonal or other changes in the snow density (for example, close-off of a high-density layer may effectively seal a deeper layer with lower density in which pores are still open internally; Raynaud and Whillans, 1982). The measurements must be corrected for ice flow. The relation between air pressure and elevation can be changed meteorologically (e.g. by a change from a consistently low-pressure to a consistently high-pressure region). Elevations and atmospheric pressures are referenced to sea level, which moves up and down significantly on glacial-interglacial time scales. In addition, we lack a physical understanding of the process of pore close-off and the observed volume-temperature relation, so we lack a strong basis for assuming that the relation has not changed over time, especially when extrapolating the relation to ice-age conditions of temperature and accumulation rate that do not exist on earth today.

The data obtained by this method are, in general, consistent with inferences from ice-flow modeling, glacial geology, and other ice-core techniques. For example, data from Camp Century in Greenland show that old ice originated at a higher altitude than the modern surface at this near-coastal site (Budd and Young, 1983). Glacial-geological evidence from along the Greenlandic coast shows that the ice sheet was advanced during the last ice age, and ice-flow modeling shows that such an advance probably would cause inland ice deposited at a higher elevation to flow through Camp Century, which now is a local center of ice outflow. In East Antarctica, data from Vostok indicate that the ice sheet was thicker during interglacials than during glacials (Raynaud and others, 1987). This is expected from ice-flow modeling, if reconstructed surface-temperature and accumulation-rate histories are accurate (Whillans, 1981).

Preliminary results from Vostok seemed to show that the ice sheet thickened more rapidly during deglaciation than the reconstructed accumulation rate (Raynaud and others, 1987), and this has been used in informal discussions (e.g. at the first SeaRISE workshop) to argue that total-gas measurements are not reliable. However, the data and interpretations are not yet published, and were not finished at the time they were presented. Moreover, the reconstructed ice-age conditions of low temperature and accumulation rate at Vostok are not duplicated on earth today, requiring an extrapolation rather than interpolation of the modern temperature-volume relation, and introducing greater uncertainty. Finally, there are real questions about how the pressure field over Antarctica varied during the last ice age, questions that should be addressed by atmospheric (or polar environmental) modeling. Pending further data, it does appear that total-gas-content measurements provide useful information on past ice-surface elevations, but further research directed to solving ice-dynamical problems clearly is warranted.

An independent approach is based on analysis of isotopic data (Grootes and Stuiver, 1986; 1987). After correction for source and path effects and ice flow, an isotopic shift in an ice core can be interpreted as a temperature change at a site. That temperature change can be decomposed into two components: a climatic change and an elevation change. If the climatic change can be inferred in some manner (e.g. by measuring the isotopic shift at a location where models or other data show that elevation changes were small, and assuming regional coherence of the climatic change), then subtracting the climatic from the total temperature change yields the change linked to elevation. By assuming a lapse rate, this elevation-linked change yields the elevation change at a site.

Clearly, there are a number of assumptions here. Particularly troubling is the observation that the relation between modern surface temperatures and elevations in West Antarctica does not follow the expected atmospheric lapse rate: Byrd Station is about 1200 m higher (Drewry, 1983), but only 1.5°C colder, than the Upstream B camp on ice stream B. (Again, the low elevation of Upstream B and the Siple Coast in general may contribute to this discrepancy.) However, the results of the method agree fairly well with results from total-gas measurements (Grootes and Stuiver, 1986; 1987).

For the West Antarctic ice sheet, detailed, multi-parameter analyses of surface elevation are available from Byrd Station only. There, it appears that the onset of deglaciation caused thickening of one to a few hundred meters (Fig. F), followed by similar thinning to the present. This is consistent with ice-flow models based on the inferred changes in accumulation rate and temperature: warming and accumulation-rate increase at the end of the ice age would have caused immediate thickening from the extra snowfall, with delayed thinning as the warmth penetrated to deeper ice and increased flow rates (Whillans, 1981; Raynaud and Whillans, 1982; Jenssen, 1983; Budd and Young, 1983; Grootes and Stuiver, 1986; 1987).

OTHER DATA.

Ice cores can provide other useful historical data. For example, if basal freeze-on has been occurring, it will be evident through a thick layer of dirty basal ice (as at Byrd; Gow and others, 1979) or of frozen-on seawater (as at J9 on the Ross Ice Shelf; Zotikov and others, 1980). The physical properties of the ice record the stress state and cumulative strain of the ice (or at least the minimum value of the cumulative strain; Alley, 1988).

Many further examples could be given. However, most would be relatively peripheral to SeaRISE, and are omitted here. Notice, however, that this in no way implies that other areas are not important science, but only that they are not central to SeaRISE goals.

SUMMARY.

For central West Antarctica, then, ice-core data and other evidence centered around Byrd Station tell a broadly consistent story. Ice-age temperatures were colder (by several degrees), accumulation was less (by perhaps a factor of two), and surface elevation was similar to modern. The end of the ice age increased temperature and accumulation rather rapidly to modern values (between perhaps 15,000 and 10,000 years ago), which caused thickening of one to a few hundred meters, followed by thinning to the present. Exact timing of events is not known accurately, however, and uncertainties in magnitudes are large. Since the end of the deglaciation, temperatures and accumulation rates probably have fluctuated little about Holocene means.

INITIAL CONDITIONS.

Ice cores will provide, and have provided, some useful information on initial conditions for SeaRISE modeling of the WAIS. This information is mostly related to mass balance.

The modern mass balance (i.e., whether the ice sheet is growing or shrinking) can be calculated if we know the ice input from snowfall and ice output from ice flow and basal melting. The input can be measured in several ways. The most successful has been collection of shallow ice cores (<20 m) with hand augers, measurement of gross beta activity to locate known reference horizons from atmospheric atomic-bomb testing (e.g. A.D. 1955), and measurement of core density; the accumulation rate averaged over the period between the reference horizon and the coring then can be calculated directly (e.g. Whillans and Bindschadler, 1988). Technically, this is ice-core analysis, although the work usually is carried out by ice dynamicists.

The ice outflow depends on the surface velocity field and the velocity-depth profile. The surface velocity field is measured by surface surveying tied to satellite tracking (TRANSIT or GPS) or through repeat imaging of specific sites (satellite or airborne photogrammetry)(SCP, 1988). The velocity-depth profile can be calculated from flow models, but should be measured for improved accuracy. Measurement is easiest through repeat inclinometry in boreholes left by ice-core recovery (Fig. G). Data from such inclinometry from the Byrd Station borehole have been used by Whillans (1977) to estimate that the ice sheet upglacier of Byrd now is thinning by a few centimeters per year.

GOVERNING EQUATIONS.

An ice-flow model requires governing equations for ice deformation, basal sliding, and subglacial deformation, linking the ice velocity or strain rate to the driving stress and other factors. For basal sliding and subglacial deformation, ice-core drilling provides access holes to the bed. Ice-core access holes are more expensive and slower to drill than hot-water holes, but remain open much longer for extended experiments, and thus are complementary to hot-water holes. (Antifreeze techniques for hot-water holes are under development, however; H. Engelhardt, pers. comm., 1990.)

Ice-core drilling is especially useful for learning the governing equations of ice deformation. The rate of ice deformation in any direction is linearly proportional to the stress deviator in that direction and to the square of the effective stress, increases exponentially with the absolute temperature, and also varies with ice fabric and dislocation substructure (Paterson, 1981, ch. 3). Ice-core drilling provides access holes in which temperature can be measured, and also allows studies to measure the ice fabric.

An ice crystal is about two or more orders of magnitude "softer" parallel to its basal plane than perpendicular to it (that is, a given shear stress oriented parallel to the basal plane will produce a shear strain rate more than 100 times larger than that produced by a shear stress of the same magnitude oriented perpendicularly to the basal plane). Deformation of a polycrystalline ice sample causes the pattern of c-axes (normals to the basal planes) to change, which changes the hardness of the ice and the deformation rate (Budd and Jacka, 1989). In natural ice, thin-section analyses supplemented by ultrasonic measurements on core samples, borehole sonic logging, and surface seismic surveys, show that a shear stress causes c-axes to cluster so that the ice is soft to that shear stress, that a small normal stress causes the ice to harden to that stress, and that a large normal stress causes the ice to soften to that stress (Alley, 1988; Budd and Jacka, 1989). In addition, softening to a given stress causes hardening to some other stresses; anisotropy of c-axis fabrics causes the ice hardness to be a tensor property. In polycrystalline ice, the difference between hard and soft to a given stress can be as large as an order of magnitude or more, as demonstrated by laboratory deformation tests on ice-core samples and by analysis of data from repeated borehole inclinometry (Budd and Jacka, 1989). Ice-flow models show that the behavior of modern ice sheets cannot be explained adequately without including effects such as these (Van der Veen and Whillans, in press).

The major results directly applicable to SeaRISE thus far are that ice subjected primarily to basal shear in inland ice (and especially in ice-age ice inland) develops a strong vertical maximum of c-axes, which softens the ice to simple shear by roughly three times (e.g. Fig. G; Dahl-Jensen and Gundestrup, 1987). Ice subjected to strong normal stresses on ice shelves forms a soft c-axis girdle pattern around the normal stresses (Budd and Jacka, 1989); seismic results suggest that this extends into ice streams as well (Blankenship and others, 1989). We can infer that as ice flows from inland into ice streams or ice shelves, it must reorganize its fabric, passing through a hard configuration (Alley, in prep.); however, this has not been tested directly. The picture in inland ice is complicated slightly by the deepest ice of the Byrd core (Gow and Williamson, 1976), which would be soft to a vertical normal stress but not to a basal shear stress; this probably is related to the unusual topographic position of Byrd near the top of a subglacial mountain.

The temperature profiles from Byrd and elsewhere in coreholes, and from Upstream B, Crary Ice Rise, and other West Antarctic sites in hot-water holes (e.g. Engelhardt and others, 1990), are useful results for flow modeling. Also, the repeat inclinometry of the Byrd hole (Hansen and others, 1989) shows how the ice deforms, which helps constrain the flow law there.

An additional useful result from ice-core analyses is the identification of marker horizons (Budd and others, 1989). An event such as a large volcanic eruption, or a spike in

^{10}Be deposition, can be identified in several ice cores (Fig. E). The depth of the event depends on accumulation rates and ice flow. Any flow model must be able to reproduce the relative depths of an event in different cores. (Internal layers detected with radar serve a similar purpose; Whillans, 1976.)

SUMMARY--AND AN EDITORIAL.

Ice-core analysis is a widely useful discipline. It provides the only direct measurement of past atmospheric compositions, and thus is a centerpiece of global-change studies. It also provides important data on local atmospheric conditions, and on behavior of the ice sheet.

Because SeaRISE studies marine ice sheets that are embedded in the global environment, the global implications of ice-core analyses are of interest to us. However, to optimize return from investment, SeaRISE needs to emphasize certain aspects of ice-core analysis. The major contributions of ice coring to SeaRISE would come (and have come) from learning the history of the ice sheet (near-surface temperatures, net surface accumulation rate, and surface elevation), the present state of the ice sheet (temperature and mass balance) and the way it flows (ice hardness). Ice cores also provide important, long-lasting access holes to the bed for ice-dynamical and geological studies.

A representative (but not exhaustive) list of the major experiments needed on ice cores to achieve these contributions includes: isotopic analyses (for past temperatures, surface elevations and accumulation rates); detailed measurements of annually varying components such as microparticles, chemistry, isotopes and visible strata (for accumulation rates); studies of cosmogenic nuclides such as ^{10}Be (for accumulation rates); total-gas-content measurements (for surface elevations); and thin-section and ice-deformation studies (for ice hardness). Related experiments in the corehole include: thermometry (ice hardness, past temperature); repeat inclinometry (mass balance, ice hardness); and sonic logging (ice hardness); as well as subglacial studies for other purposes.

For local signals, core locations are critical to answering major questions. In West Antarctica, the past configuration of the ice sheet is of special interest: Did the ice sheet collapse during the previous interglacial? Did its margins thicken and its grounding line advance during the most recent glacial? Is it collapsing now? Discovery of West Antarctic ice from the previous interglacial would show that the ice sheet existed then. The best chance of finding such ice is in regions without basal melting, as indicated by ice-flow models. If such ice is absent, measurement of geothermal flux in subglacial bedrock combined with ice-flow modeling can be used to determine whether it melted off recently or whether it was not present because of ice-sheet collapse. Inland ice is only weakly sensitive to changes at the ice-sheet margins (by analogy to Alley and Whillans, 1984), so data from marginal regions are needed to learn what has happened at the margins during deglaciation. For example, total-gas and isotopic data from a core only a few hundred meters long on ridge BC on the Siple Coast would reveal whether the ice-age ice sheet there was significantly thicker than today (Graham, 1988).

In the U.S., ice-core analysis often is equated with ice-core geochemistry. In the case of the Greenland Ice Sheet Project 2 (GISP2) drilling, borehole measurements, ice-core geophysics and surface glaciology were discouraged purposefully, to concentrate on ice-core geochemistry. This was entirely appropriate, given the goals of the project, its limited funding, and the possibility of adding-on studies during or after coring. However, to optimize the value of future ice coring to SeaRISE, it is imperative that GISP2 not be used as a model. Rather, glaciological as well as geochemical aspects should figure in site selection, and the geochemical studies must be coupled to a full suite of "glaciological" measurements: total gas content for surface elevations, ice-core physical properties, borehole logging for sonic velocity, temperature, and inclination change, surface surveys of accumulation, temperature, and ice movement, surface geophysical surveys, etc. This would ensure that information on both global and local conditions would be obtained.

In addition, there are a number of opportunities for interdisciplinary research that would greatly improve the scientific return from ice-core analyses. The major openings seem to be in the fields of atmospheric modeling and ice-flow modeling. Models should be formulated to track features that can be measured in ice cores (e.g. isotopes for atmospheric modeling, ice temperatures and the depth of marker horizons for ice flow). Joint analysis of ice-core, atmospheric and ice-flow data then can improve the interpretation of the ice cores and of the dynamics of the entire system.

Few truly solid results have been obtained from ice-core studies that are directly relevant to SeaRISE. However, no major effort has been mounted to obtain such results for more than twenty years. The many techniques available to us provide considerable confidence that a well-developed ice-coring program would provide much information absolutely essential to the success of SeaRISE.

Acknowledgements. I thank P. Grootes and I. Whillans for good ideas, R. Bindschadler for impetus, and NSF for funding to study ice cores.

References.

Alley, R.B. In preparation. Flow-law hypotheses for ice-sheet modeling. For *Journal of Glaciology*; available from author on request.

Alley, R.B. 1988. Fabrics in polar ice sheets: development and prediction. *Science* **240**(4851), 493-495.

Alley, R.B. and C.R. Bentley. 1988. Ice-core analysis on the Siple Coast of West Antarctica. *Annals of Glaciology* **11**, 1-7.

Alley, R.B. and B.R. Koci. 1990. Recent warming in central Greenland? *Annals of Glaciology* **14**, 5-8.

Alley, R.B. and I.M. Whillans. 1984. Response of the East Antarctic ice sheet to sea-level rise. *Journal of Geophysical Research* **89**(C4), 6487-6493.

Benoist, J.P., J. Jouzel, C. Lorius, L. Merlivat and M. Pourchet. 1982. Isotope climatic record over the last 2.5 ka from Dome C, Antarctica, ice cores. *Annals of Glaciology* **3**, 17-22.

Blankenship, D.D., R.B. Alley and C.R. Bentley. 1989. Fabric development in ice sheets:

seismic anisotropy. [Abstract]. *Eos* **70**(15), 462.

Bolzan, J.F. 1985. Ice flow at the Dome C ice divide based on a deep temperature profile. *Journal of Geophysical Research* **90**(D5), 8111-8124.

Bromwich, D.H. 1988. Snowfall in high southern latitudes. *Reviews of Geophysics* **26**(1), 149-168.

Budd, W.F. and T.H. Jacka. 1989. A review of ice rheology for ice sheet modelling. *Cold Regions Science and Technology* **16**, 107-144.

Budd, W.F. and N.W. Young. 1983. Application of modelling techniques to measured profiles of temperatures and isotopes. In *The Climatic Record in Polar Ice Sheets*, G. de Q. Robin, ed., Cambridge University Press, Cambridge, p. 150-177.

Budd, W.F. and 11 others. 1989. How can an ice core chronology be established? In *The Environmental Record in Glaciers and Ice Sheets*, H. Oeschger and C.C. Langway, Jr., eds, Wiley, New York, p. 177-192.

Chappellaz, J., J.M. Barnola, D. Raynaud, Y.S. Korotkevich and C. Lorius. 1990. Ice-core record of atmospheric methane over the past 160,000 years. *Nature* **345**(6271), 127-131.

Clausen, H.B., N.S. Gundestrup, S.J. Johnsen, R. Bindschadler and H.J. Zwally. 1988. Glaciological Glaciological investigations in the Crete area, central Greenland: a search for a new deep-drilling site. *Annals of Glaciology* **8**, 10-15.

Dahl-Jensen, D. and N.S. Gundestrup. 1987. Constitutive properties of ice at Dye 3, Greenland. *IAHS Publ. No. 170*, 31-43.

Dahl-Jensen, D. and S.J. Johnsen. 1986. Paleotemperatures still exist in the Greenland ice sheet. *Nature* **320**, 250-252.

Dansgaard, W. and H. Oeschger. 1989. Past environmental long-term records from the Arctic. In *The Environmental Record in Glaciers and Ice Sheets*, H. Oeschger and C.C. Langway, Jr., eds, Wiley, New York, p. 287-318.

Drewry, D.J., ed. 1983. *Antarctica: Glaciological and Geophysical Folio*. Scott Polar Research Institute, Cambridge.

Engelhardt, H., N. Humphrey, B. Kamb and M. Fahnestock. 1990. Physical conditions at the base of a fast moving Antarctic ice stream. *Science* **248**, 57-59.

Gow, A.J. 1968. Deep core studies of the accumulation and densification of snow at Byrd Station and Little America V, Antarctica. *U.S. Army CRREL Research Report 197*.

Gow, A.J. and T. Williamson. 1976. Rheological implications of the internal structure and crystal fabrics of the West Antarctic ice sheet as revealed by deep core drilling at Byrd Station. *U.S. Army CRREL Report 76-35*.

Gow, A.J., S. Epstein and W. Sheehy. 1979. On the origin of stratified debris in ice cores from the bottom of the Antarctic ice sheet. *Journal of Glaciology* **23**(89), 185-192.

Graham, R.F. 1988. Deep ice core site evaluation for ridge B/C, West Antarctica. B.Sc. Thesis, Dept. of Geology and Mineralogy, The Ohio State University, Columbus.

Grootes, P.M., M. Stuiver, T.L. Saling, P.A. Mayewski, M.J. Spencer, R.B. Alley and D. Jenssen. 1990. A 1400-year oxygen isotope history from the Ross Sea area, Antarctica. *Annals of Glaciology* **14**, 94-98.

Grootes, P.M. and M. Stuiver. 1986. Ross Ice Shelf oxygen isotopes and West Antarctic climate history. *Quaternary Research* **26**, 49-67.

Grootes, P.M. and M. Stuiver. 1987. Ice sheet elevation changes from isotope profiles. *IAHS Publ. No. 170*, 269-281.

Hammer, C.U. 1989. Dating by physical and chemical seasonal variations and reference horizons. In *The Environmental Record in Glaciers and Ice Sheets*, H. Oeschger and C.C. Langway, Jr., eds, Wiley, New York, p. 99-121.

Hammer, C.U., H.B. Clausen and C.C. Langway, Jr. 1985. The Byrd ice core: continuous acidity measurements and solid electrical conductivity measurements. [Abstract]. *Annals of Glaciology* **7**, 214.

Hansen, B.L., J.R. Kelty and N.S. Gundestrup. 1989. Resurvey of Byrd Station drill hole, Antarctica. *Cold Regions Science and Technology* **17**(1), 1-6.

Harvey, L.D.D. 1988. Climatic impact of ice-age aerosols. *Nature* **334**(6180), 333-335.

Jenssen, D. 1983. Elevation and climatic changes from total gas content and stable isotopic measurements. In *The Climatic Record in Polar Ice Sheets*, G. de Q. Robin, ed., Cambridge University Press, Cambridge, p. 138-144.

Johnsen, S.J. 1977. Stable isotope profiles compared with temperature profiles in firn with historical temperature records. *IAHS Publ. No. 118*, 388-392.

Johnsen, S.J. and G. de Q. Robin. 1983. Diffusion of stable isotopes. In *The Climatic Record in Polar Ice Sheets*, G. de Q. Robin, ed., Cambridge University Press, Cambridge, p. 57-63.

Jouzel, J., L. Merlivat and C. Lorius. 1982. Deuterium excess in an East Antarctic ice core suggests higher relative humidity at the oceanic surface during the last glacial maximum. *Nature* **299**, 688-691.

Jouzel, J., G.L. Russell, R.J. Suozzo, R.D. Koster, J.W.C. White and W.S. Broecker. 1987. Simulations of the HDO and $H_2{}^{18}O$ atmospheric cycles using the NASA/GISS general circulation model: the seasonal cycle for present day conditions. *Journal of Geophysical Research* **92**, 14,739-14,760.

Jouzel, J. and 9 others. 1989. A comparison of deep Antarctic ice cores and their implications for climate between 65,000 and 15,000 years ago. *Quaternary Research* **31**, 135-150.

Lorius, C. 1983. Antarctica: survey of near-surface mean isotopic values. In *The Climatic Record in Polar Ice Sheets*, G. de Q. Robin, ed., Cambridge University Press, Cambridge, p. 52-56.

Lorius, C., G. Raisbeck, J. Jouzel and D. Raynaud. 1989. Long-term environmental records from Antarctic ice cores. In *The Environmental Record in Glaciers and Ice Sheets*, H. Oeschger and C.C. Langway, Jr., eds, Wiley, New York, p. 343-361.

MacAyeal, D., E. Waddington and J. Firestone. In review. Paleothermometry by control methods. *Journal of Glaciology*.

Mosley-Thompson, E. and L.G. Thompson. 1982. Nine centuries of microparticle deposition at the South Pole. *Quaternary Research* **17**(1), 1-13.

Mosley-Thompson, E., L.G. Thompson, P.M. Grootes and N. Gundestrup. 1990. Little Ice Age (Neoglacial) paleoenvironmental conditions at Siple Station, Antarctica. *Annals of Glaciology* **14**, 199-204.

Paterson, W.S.B. 1981. *The Physics of Glaciers, Second Edition*. Pergamon, Oxford.

Peel, D.A., R. Mulvaney and B.M. Davison. 1988. Stable-isotope/air-temperature relationships in ice cores from Dolleman Island and the Palmer Land Plateau, Antarctic Peninsula. *Annals of Glaciology* **10**, 130-136.

Raisbeck, G.M., F. Yiou, D. Bourles, C. Lorius, J. Jouzel and N.I. Barkov. 1987. Evidence for two intervals of enhanced ^{10}Be deposition in Antarctic ice during the last glaciation period. *Nature* **326**, 273-277.

Raynaud, D. 1983. Total gas content. In *The Climatic Record in Polar Ice Sheets*, G. de Q. Robin, ed., Cambridge University Press, Cambridge, p. 79-82.

Raynaud, D. and I.M. Whillans. 1982. Air content of the Byrd core and past changes in the West Antarctic ice sheet. *Annals of Glaciology* **3**, 269-273.

Raynaud, D., D. Mazaudier, C. Lorius, N.I. Barkov and V. Lipenkov. 1987. Elevation changes over the past 160 000 years near Vostok, East Antarctica from air content in ice. [Abstract]. *IAHS Publ. No. 170*, 297.

Robin, G. de Q. 1977. Ice cores and climatic change. *Philosophical Transactions of the Royal Society of London, Series B* **280**, 143-168.

Robin, G. de Q., 1983. Profile data, inland Antarctica. In *The Climatic Record in Polar Ice Sheets*, G. de Q. Robin, ed., Cambridge University Press, Cambridge, p. 112-118.

Robin, G. de Q. and S.J. Johnsen. 1983. Atmospheric processes. In *The Climatic Record* *in Polar Ice Sheets*, G. de Q. Robin, ed., Cambridge University Press, Cambridge, p. 47-52.

SCP (Siple Coast Project Steering Committee). 1988. *Science plan for the Siple Coast Project.* Byrd Polar Research Center, The Ohio State University, Columbus.

Shackleton, N.J. 1983. Isotopic composition of the ocean surface as a source for ice in Antarctica. In *The Climatic Record in Polar Ice Sheets*, G. de Q. Robin, ed., Cambridge University Press, Cambridge, p. 43-47.

Staffelbach, T., B. Stauffer and H. Oeschger. 1988. A detailed analysis of the rapid changes in ice-core parameters during the last ice age. *Annals of Glaciology* **10**, 167-170.

Thompson, L.G., E. Mosley-Thompson and J.R. Petit. 1981. Glaciological interpretation of microparticle concentrations from the French 905-m Dome C, Antarctica core. *IAHS Publ. No. 131*, 227-234.

Van der Veen, C.J. and I.M. Whillans. In press. Flow laws for glacier ice: comparison of numerical predictions and field measurements. *Journal of Glaciology.*

Whillans, I.M. 1976. Radio-echo layers and the recent stability of the West Antarctic ice sheet. *Nature* **264**, 152-155.

Whillans, I.M. 1977. The equation of continuity and its application to the ice sheet near 'Byrd' Station, Antarctica. *Journal of Glaciology* **18**, 359-371.

Whillans, I.M. 1981. Reaction of the accumulation zone portions of glaciers to climatic change. *Journal of Geophysical Research* **92**(C5), 4274-4282.

Whillans, I.M. and R.A. Bindschadler. 1988. Mass balance of ice stream B, West Antarctica. *Annals of Glaciology* **11**, 187-193.

Zotikov, I.A., V.S. Zagorodnov, and J.V. Raikovsky. 1980. Core drilling through the Ross Ice Shelf (Antarctica) confirmed basal freezing. *Science* **207**, 1463-1465.

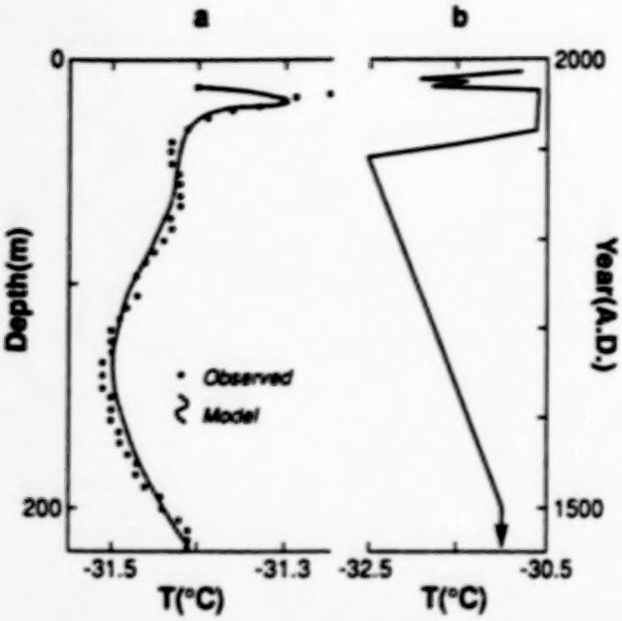


Fig. 1a. Temperature-depth profile,* in °C. **Data** (·) measured with first thermistor (second thermistor was consistently 0.035°C colder), and model results (smooth curve). **Seasonal** effects from the two years prior to drilling are evident between 12 and 15 m: they are large (but not plotted) shallower than 12 m. 1b. **Best**-fit history of surface temperature.

Fig. A. Temperature data and interpretation, from GISP2, Greenland (Alley and Koci, 1990).

Figure 5.53. Oxygen isotopic ratios for Byrd Station (after Epstein *et al.*, 1970) compared with curve calculated using flowline model.

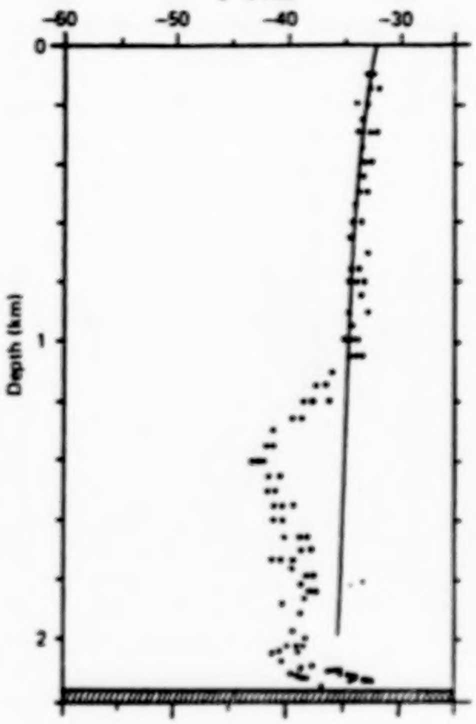


Fig. B. Isotopic data and correction for flow, Byrd Station, Antarctica (Budd and Young, 1983).

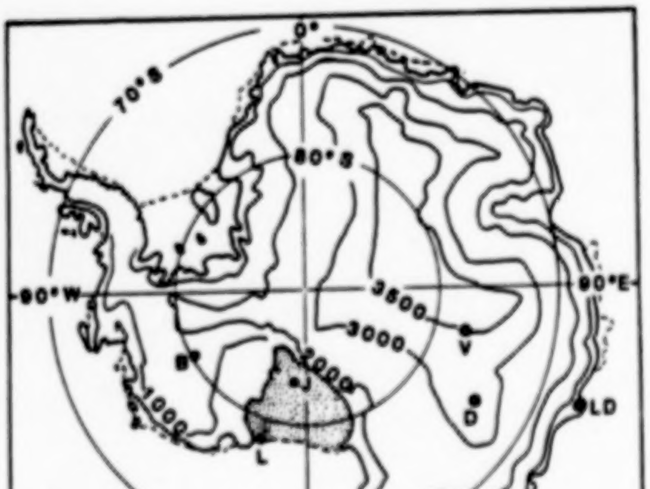


FIG. 1 Localities of Antarctic cores discussed. B: Byrd Station; D: Dome C; D-10: Terre Adélie, core D-10; L: Little America V; LD: Law Dome, cores BHF, BHC-1, BHC-2; J: J-9; V: Vostok.

Fig. C. Locations of Antarctic cores (Grootes and Stuiver, 1987).

Figure 3.11. Mean near-surface δ values plotted against mean air temperature at ground level. Key to symbols shown on diagram. δ values are from Lorius *et al.* (1969, Table 1) and from the following: Dansgaard (unpublished?) for D45, D59, D80; Gonfiantini (1965) for Roi Baudouin; Epstein, Sharp & Goddard (1963) for Little America, Wilkes S2; Dansgaard *et al.* (1973) for Horlick Mountains, Halley Bay and Byrd; Vilenskiy *et al.* (1970) for Mirny; Vilenskiy *et al.* (1974) for Molodezhnaya and Vostok δ ^{18}O values; Epstein & Sharp (1967) for Eights; Aldaz & Deutsch (1967) for South Pole; Picciotto (1967) for Plateau Station; Picciotto *et al.* (1968) for Pole of Relative Inaccessibility; measurements by Lorius (unpublished) on samples from Wolmarans from SA9 and Peel from HB3. Line 1 for inland stations from Lorius & Merlivat (1977), line 2 for coastal stations from data in Budd & Morgan (1973). Points in Area 3 are from Marie Byrd Land and Area 4 from coastal stations. Some temperature values are from Bentley *et al.* (1964).

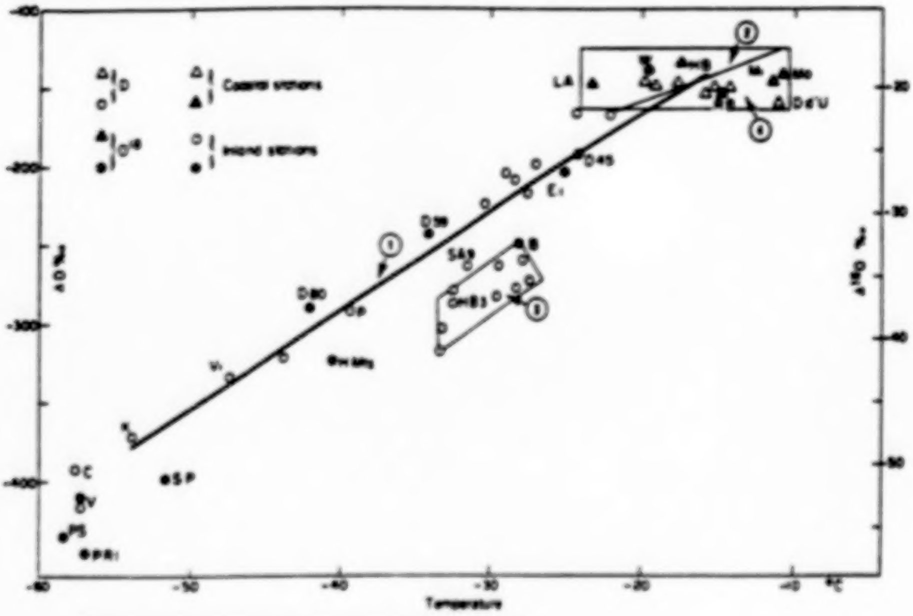


Fig. D. Variation of surface isotopic values in Antarctica (Lorius, 1983). Note anomalous W. Antarctic data.

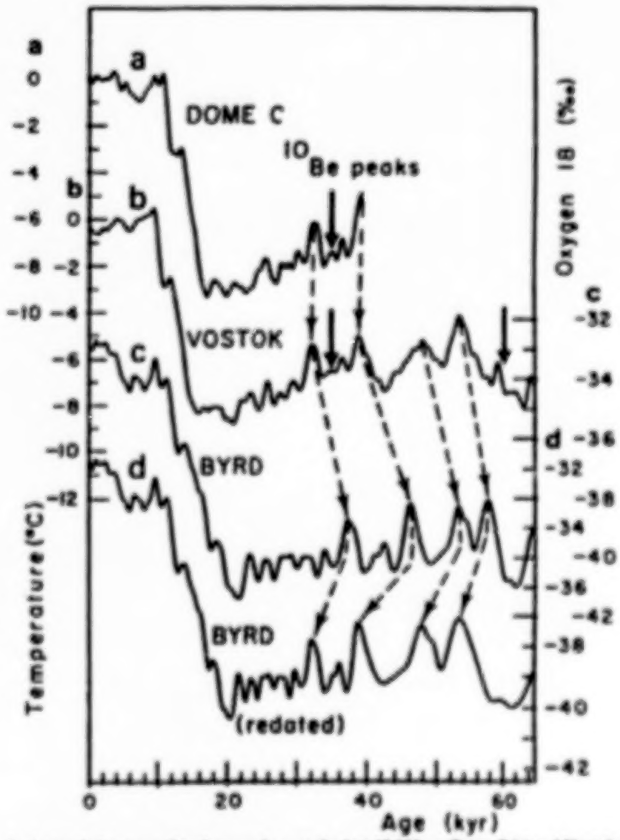


Fig. E. Antarctic-core isotopes and interpreted temperatures, with core correlations based on ^{10}Be spikes (Jouzel and others, 1989).

FIG. 3. Antarctic ice core climatic records over the last 65,000 yr. Dome C (a) and Vostok (b): left scale in °C expressed as a difference with respect to current surface temperature value. Solid arrows indicate observed ^{10}Be peaks. Byrd (c): right scale in δ^{18}O‰ (from Johnsen *et al.*, 1972); time scale from H. B. Clausen and C. U. Hammer (personal communication, 1988). The dotted arrows suggest a possible correspondence of climatic events in the three cores. Curve d, the Byrd record, uses modified time scale obtained using these events as time markers.

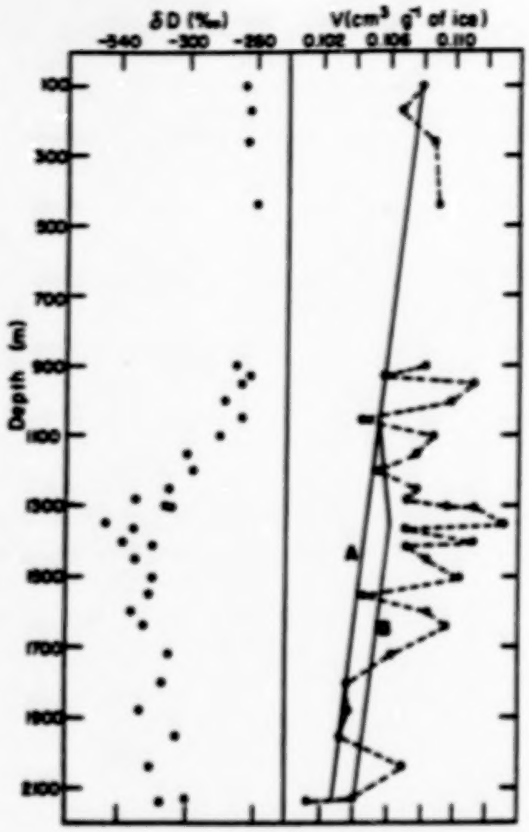


Fig.3. Byrd station. Variations of the mean total gas content V and isotope ratio δ with depth. Mean V values measured at different depth levels are compared with the change of V to be expected by ice flow according to the present flow pattern under constant temperature (line A) and corrected for colder temperatures during the ice age (curve B).

Fig. F. Total-gas-content data at Byrd Station, showing decrease at end of ice age probably caused by ice thickening (Raynaud and Whillans, 1982).

Fig. G. Borehole and core data from Dye 3, Greenland (Dahl-Jensen and Gundestrup, 1987). Borehole inclinometry shows enhanced softness of ice-age ice (which has high dust content and small crystals).

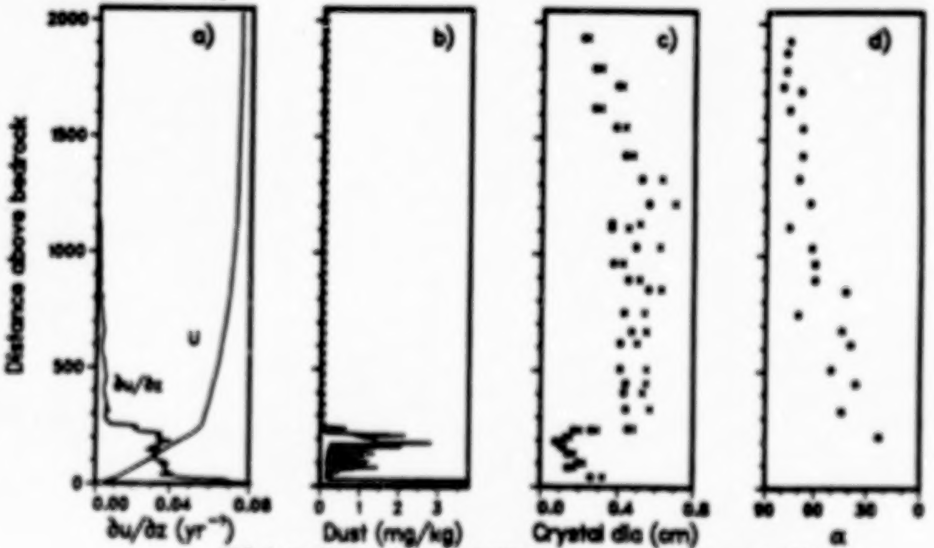


FIG. 2 (a) Profiles of horizontal velocities and shear strain rate derived from the inclinometer surveys of the Dye 3 borehole (Fig. 2a). The bulk profile can be divided into three zones: the bottom 25 m of silty ice with very high deformation rates; the immediately overlying 230 m of ice, deposited during the last glaciation; and the uppermost 1780 m of Holocene ice, with lower deformation rates. (b) Dust concentration, the bulk concentration of insoluble microparticles in the ice. (Hammer et al., 1985). In the silty ice the dust concentration is > 20 mg/kg. (c) Crystal diameter (Herron et al., 1985). The circles represent vertical dimensions and the x's represent the horizontal dimensions. (d) Fabric parameter α, the half-apex angles of the cones containing 90% of the c-axis (Herron et al., 1985).

GLACIAL DYNAMICS (glaciology)

Ian Whillans[1]

I. Introduction

This report reviews recent results from studies of ice dynamics that relate to the objectives of the WAIS initiative. It emphasizes what we have discovered, which is very considerable. It also emphasizes what we do not understand in order to stimulate further discussion.

The best evidence shows that the ice sheet in West Antarctica is the most rapidly changing ice sheet on earth today. Its rate of change is much faster than most glaciologists had expected and it is changing in a manner much more complex than foreseen.

Follow-on questions are 'why is the ice sheet changing and why so rapidly ?'. The changes have two broad causes:

(1) a delayed but ongoing response to the termination of the last glaciation about 10 000 years ago

(2) automatic, internally-caused flow adjustments.

It is not fully known why the response to the last global termination is so delayed, nor is the operation of internal instabilities understood, and certainly we are not yet in a position to predict the future course of the evolution of the ice sheet. Deeper study of special features of the West Antarctic Ice Sheet is needed. Many of these studies require multidisciplinary approaches.

II. Styles of Ice Flow

After the discovery of the mass imbalances, the most stunning discovery from the studies of the ice streams is that there are several styles of ice flow. Very likely these two discoveries are related, in that the mass imbalances result from switches between styles of flow. The flow styles are listed below in order of slowest to fastest speed. Examples are labelled in Figure 1.

---

[1] Byrd Polar Research Center and Department of Geological Sciences, 125 South Oval Mall, Columbus OH 43210

'Smooth' Ice Flow

Where deformation rate at depth is slow, the most important deformation style is vertical compression and horizontal spreading. With time and net snow accumulation, the glacier adjusts to form a smooth upper surface.

Smooth ice is found near ice divides and on ice shelves. Most of the interstream ridges in West Antarctica are smooth.

'Mottled' Ice Flow.

Where deep ice speed is faster and there are basal irregularities, the ice surface is alternately steeper and flatter, forming steps or undulations. If the bed variations are random, so are the disturbances to the surface, and a mottled pattern is evident in satellite imagery. Often the mottles appear to be aligned in a direction perpendicular to solar illumination. One should not be misled by this solar bias.

Mottled surfaces dominate for much of the grounded inland ice. Parts of the interstream ridges are also mottled (for example, ridge A/B). This could be due to locally large bed variations or more likely to locally (and anomalously) faster ice flow.

'Streaming' Ice Flow.

As ice streams form, the mottles give way to linear stripes variously called streamlines, flowlines, sutures, septa, plumes, and flow traces. In their extreme development, they are reminiscent of medial moraines on a valley glacier, although, of course, they contain no rock debris and occur in a zone of net snow accumulation.

The leading contenders for the genesis of flow traces are (1) that they are a plume of buried crevasses, and the traces stand high because of lower mean density, and (2) that they are made of softer ice which is squeezed upward by lateral compression, the ice is soft because of different crystal size or orientation.

The courses of ice streams are influenced by bed topography and their surface is lower than that of the inland ice to either side. Their lateral boundaries are strong shear zones.

The ice streams can be extremely fast (> 800 m/a) and are restrained mainly by basal friction. The basal friction is small because of effective basal lubrication. The leading contenders for the lubricant are:

(1) continuous water film,
(2) discontinuous water film punctuated with 'sticky spots',
(3) water and debris slurry with sticky spots, and
(4) a continuous very viscous debris-with-water paste.

The development of an understanding of this lubrication should constitute a major effort within the WAIS initiative.

There are also interesting problems associated with the boundaries to the ice streams, none of which appear to be stable. The lateral boundaries are easily mapped by recognizing the shear margins. The upper and lower boundaries are not so distinct. Switches in flow style from inland ice, smooth or mottled, to streaming sometimes occurs in jumps, and chunks of inland ice are incorporated into the ice stream and carried down-flow as rafts. The process is not understood. Furthermore, there seem to be examples of margin migration due to narrowing of ice streams.

These hyperactive ice streams are potential means for the rapid evacuation of the inland-ice reservoir. Indeed, it is entirely possible that such an evacuation is underway now.

Ice Shelf Flow.

The floating portion of the ice sheet is called ice shelf and is controlled by drag past grounded ice or rock. Early theory advocated great importance to backstress (dynamic forces from the ice shelf on grounded ice) but where measured this has been relatively small. However, being part of the train of flow from ice divide to the calving front, any change in the ice shelf must propagate up-flow. The propagation time for East Antarctica is calculated to be thousands of years (Figure 2), but is probably in the range of hundreds of years for the lower-elevation portions of West Antarctica.

Major interest in the ice shelves stems from their thinness, coastal position, and relation with ocean currents:

- their thinness makes them more quickly reactive to climatic change because the ice shelf's total strength is rapidly affected by changes in ice temperature or surface mass balances.

- their coastal position and high snowfall means that quite small changes in atmospheric circulation and the delivery of snow can make large early impacts on the ice sheet in this region.

- a major component in the mass balance is basal melting and freezing. This is largely governed by ocean dynamics. A major objective of the Filchner-Ronne-Ice-Shelf-Programme (FRISP) is concerned with this. A very strong coupling between ocean and ice-shelf dynamics has been documented, and major generation of Antarctic Bottom Water is proved.

The ice shelf also contains flow traces. These are distorted because of raft incorporation, and also perhaps, fluctuations in ice stream discharge. The result is that the position of the flow traces at the calving front changes with time. This may be very important to the control of calving rate, and it may be noted that the recent breakout from the Filchner ice shelf followed such sutures. There are also enormous crevasses on the Ross Ice Shelf, well back from the barrier. The stability of the barrier has not been studied. Weakness due to flow traces and crevasses may be important.

III. Input

Probably the most important single controlling process for ice sheets is the net accumulation rate. It's broad pattern today in East Antarctica is relatively simple; nearly all of it falls near the coast. In West Antarctica it is more complex (Figure 3), it is larger near the seaward coast but is about 150 mm of ice per year around the West Antarctic ice streams. Within the region of the ice streams it is about 100 mm/year. The cause of this hole is not known. The depth of the hole is about the same, in relative terms, as the negative mass balance of ice stream B.

The value and distribution of net accumulation, for both today and in the past, is crucial to the ice sheet. It is the primary reason for the existence of the ice sheet, yet we have a very poor understanding of it.

The accumulation rate has been analysed by several disciplines. Glaciologists have taken the approach of mapping the present-day distribution and attempting to infer the causal processes from this map. This approach has been able to account for only the coarsest scale of variation. Meteorologists have analysed radiosonde data and satellite imagery, and have had some success in East Antarctica (D. Bromwich), but there has been little modern work for West Antarctica. Past changes have been inferred from ice core work, in particular for the very dry interior at Vostok, but the inferences, if valid there, can be extended only very conjecturally to sites at lower and snowier elevations. Glaciologists are generally not equipped to address this problem, and we urge sister scientists to take up this question.

IV. Relative Sea Level

This is the other most crucial control on ice sheets, especially for East Antarctica and much of Greenland, and most especially for the marine ice sheet of West Antarctica. The response to changes in sea level is quite rapid (Figure 2). Only just recently have good eustatic sea level curves become available for the past 20 000 years. These can be used to calculate relative sea level changes for various portions of Antarctica. With the emerging understanding of ice-shelf and ice-stream mechanics, and with the possibility to measure off-shore gravity and crustal rebound rates, there would seem to be good prospects for meaningfully modelling this effect, provided there are reliable data on past ice thickness against which to check the results.

V. Surface Temperature

This is important to the ice sheet because (1) it affects the stiffness of ice, (2) it is a factor determining whether the glacier can slip at its bed, and the rate of basal melt or freeze, and (3) it determines the possibility of penetrating melt water. Some authors also argue that the surface accumulation rate is simply linked with temperature, but that is debatable, especially for low-elevation or coastal sites.

Surface temperature is less important to the objectives of the WAIS initiative than some of the other factors. It is more important at longer time scales. Changes in surface temperature take a long time to penetrate the ice sheet (Figure 4). Most of the inland ice and the ice streams are wet-based and would remain so even with quite large changes in surface temperature. Only exceedingly dramatic and rather improbable climatic changes could start significant meltwater penetration in Antarctica.

VI. Mass Balance Under Ice Shelves

This affects the mechanics of the ice sheet by (1) altering its mass, thickness and extent, and (2) changing the stiffness of the ice shelf.

VII. Mass Balance Under Grounded Ice

Under ordinary circumstances, this is relatively unimportant as a cause of glacial variation. Normal geothermal plus ice-frictional heat melts some 10 to 20 mm of ice per year. This is small compared with surface mass balance, but is adequate and important in providing water-lubrication to sliding.

Landsat imagery shows a large feature of positive relief with a half-moon shape, in the center of ice stream E (Figure 5) that looks like a large sub-glacially formed volcano that is little eroded. Such an eruption can have dramatic effects on the ice sheet. Maybe the volcano is new and is affecting the ice sheet now.

VIII. Changes During the Past 20 000 Years

This history is important to understanding the present behavior of the ice sheet. The evidence comes from several lines:

(a) results from the Byrd Station core hole and radio-echo layers up-flow from Byrd Station indicate that the ice sheet is now near its maximum size and was thinner during the Wisconsinan-equivalent. A problem is that the scatter in the total gas measurements on the core is severe.

(b) relict flow features in the existing ice sheet are interpreted to show changes in flow on shorter time scales.

(c) modern mass balance show rapid changes

(d) the interpretation of the glacial-geologic record and of the presence of concentrations of meteorites on blue ice leads to the following conclusions:

- at 20 000 years BP, the interior regions of East Antarctica were similar, or perhaps (in accord with the interpretation of the Vostok ice core) a little thinner than today.

- at 20 000 years BP, much or most of the present region of the Ross Ice Shelf was grounded.

- at 20 000 years BP, the region around McMurdo Sound contained nearly stagnant ice. There were trapped lakes and the ablation rate was only on the order of 0.001 m/a [2].

---

[2] Calculated from Denton and others, 1989, , *Quaternary Research*, **31**, p. 157, fig. 5. At 167°E, the surface slope is 100 m / 15 km, and ice thickness is greater than 600 m. Thus the driving stress is > 36 kPa. Using a stiffness parameter of 500 kPa $a^{1/3}$, which corresponds to cold ice, and n = 3, and assuming lamellar flow, the mean ice velocity is $[2/(n+2)](tau/B)^n H$, in the common notation. This works out to be 0.1 m/a, a very small number, plus any basal slip. Selecting a different stiffness could increase this perhaps ten times. The down-glacial flowline is 70 km long with little lateral divergence, and so for balance, and an ice flux of 0.1 m/a, the ablation rate must be about 1 mm/a, or somewhat more if there was basal sliding or the ice were

- a major retreat of ice occurred from the McMurdo Sound region, ending about 6 000 years BP

(e) results from the marine record are not as clear. Some kind of till was deposited over most of the continental shelf of the Ross Sea, and there are also deposits in deeper water.

IX. Ice Flow Mechanics

Here is a brief summary of the field:

- Internal deformation can be modelled relatively well. It is important to the flow of inland ice, where frozen to the bed, as for example, in Greenland, and of ice shelves. There are discrepancies between theory and observation, but these are minor compared to the problems in understanding basal slip.

- Basal motion is very poorly understood. The ice streams demonstrate an interesting paradox. Often basal slip is supposed to increase as basal drag increases. The ice streams demonstrate the reverse, fast slip under small stress. There must be much subglacial water or mud, or the bed must be smooth. The debate on this has been enlivened by the discovery of a soft bed under ice stream B, that may itself be deforming.

- Dimensions of ice streams. There is no known control on the width or upglacial extents of the ice streams.

- Interplay between erosion and deposition and ice-sheet evolution. Especially at the grounding line, it is possible for the ice sheet to deposit its own bed, or erode it easily.

- Calving. This has been largely ignored. No good theory ex

X. Subglacial Geology

The style of ice flow changes where the basal relief changes. V robably the boundary is between a hard bed, with relief determined by geologic stru e, and a soft bed of tilted marine clays, with relief formed partly by glacial processes.

---

softer. The calculations could be made more carefully, but, by any account, the ablation rate in this region was very small.

XI. Ice Flow Modelling

The results of this can be only as good as the understanding of the physics governing ice flow, as noted above. An all-inclusive model is not yet appropriate, but models to test and evaluate certain aspects have been and are very helpful.

XII. The WAIS Initiative

The Siple Coast Project has discovered that the ice sheet is changing rapidly now, and that it has features that seem to make it prone to rapid and massive change. The implications have first-order global significance.

An attack on this problem requires much more than just glaciologists. For example, some of the not-fully glaciological questions that need solution are:

- What are the controls on surface mass balance ? ... role of sea ice, katabatic flow, etc..

- What has been the time-variation in surface mass balance, especially at low-elevations?

- What has been the history of change of the glaciers, especially during the past 20 000 years, and for prior intervals of glacial fluctuation ? Representative sites need to be studied, as well as arid sites such as the dry valleys.

- What are the values, and what controls the rate of melt / freeze at the base of the ice shelf?

- Is there a possibility of active subglacial volcanism in West Antarctica?

There are, no doubt, many other questions, and some of those above will need revision, but maybe this list can stimulate discussion of the problems and research strategy for The WAIS initiative.

The glaciologic discoveries in West Antarctica have been very considerable. They have lead to a quantum leap in the science of glaciology, in which the relative dominance of theory and observation have been reversed. We have emerged from a time of numerous untested theories, to a time with new data that contradict many older theories and a shortage of good theory on the crucial processes.

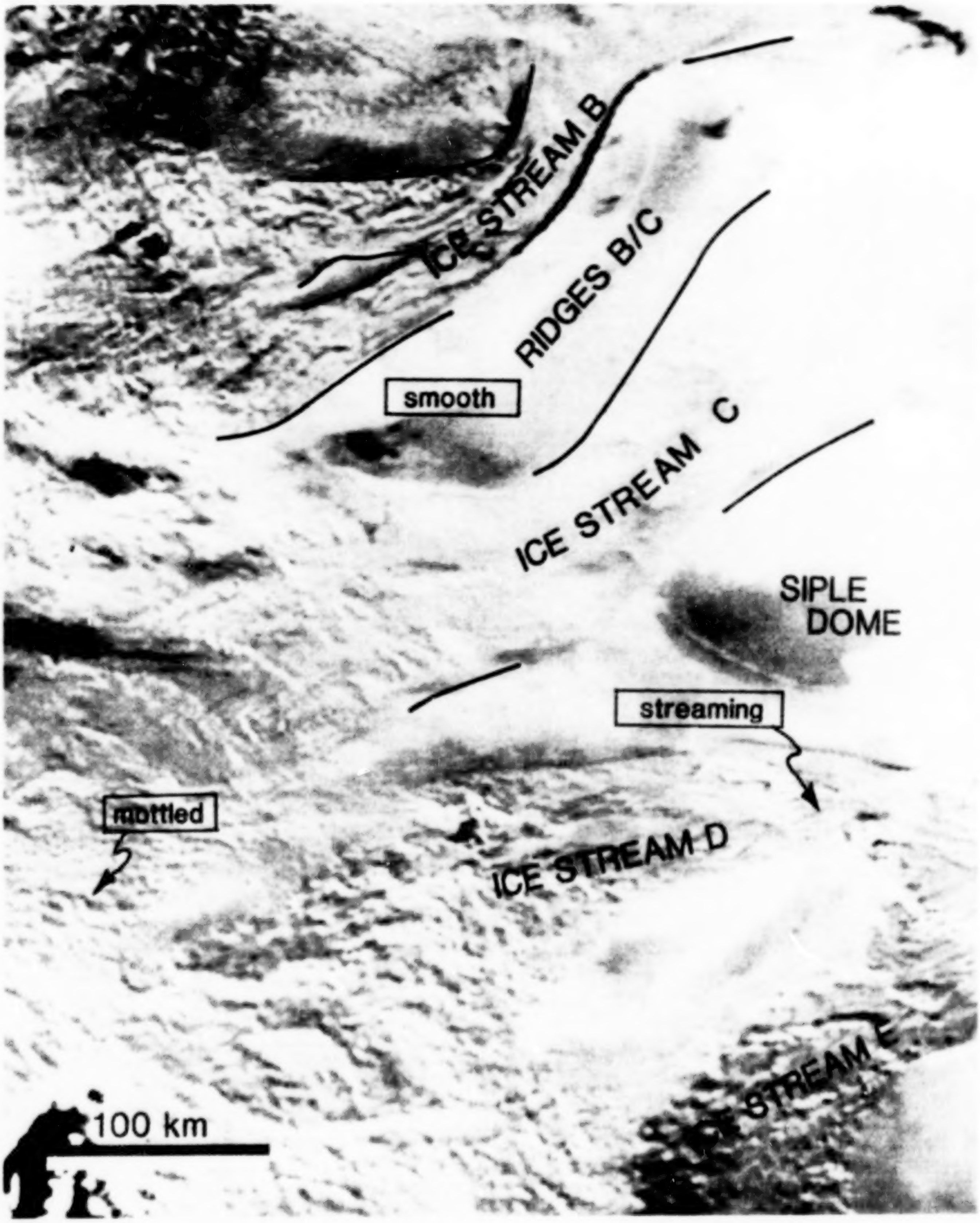


Figure 1. Satellite image of ice streams B, C, D, and E and vicinity. Flow is left to right. The Ross Ice Shelf is off-scene to the right. South is toward the top, and solar illumination is from the right. Image obtained by AVHRR (R.A. Bindschadler and P.L. Vornberger, EOS, **71**(23), 1990 cover.

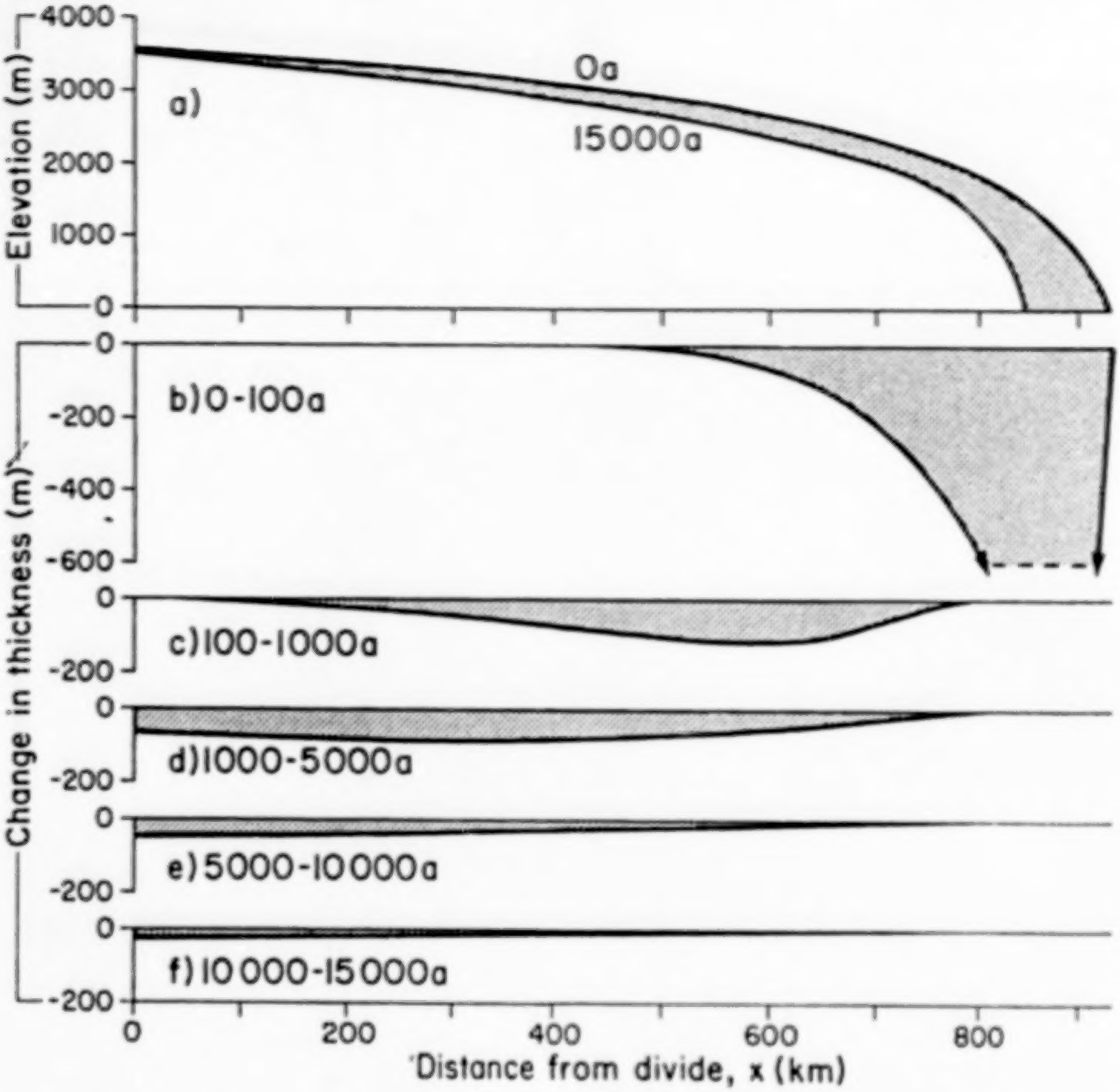


Figure 2. Progression of adjustment to a change in sea level for the East Antarctic Ice Sheet. In this calculation, the adjustment is controlled by internal ice deformation. It is mainly complete after 5000 years. Most glaciologists expect a much faster response for the ice sheet in West Antarctica. Figure from Alley and Whillans, JGR.

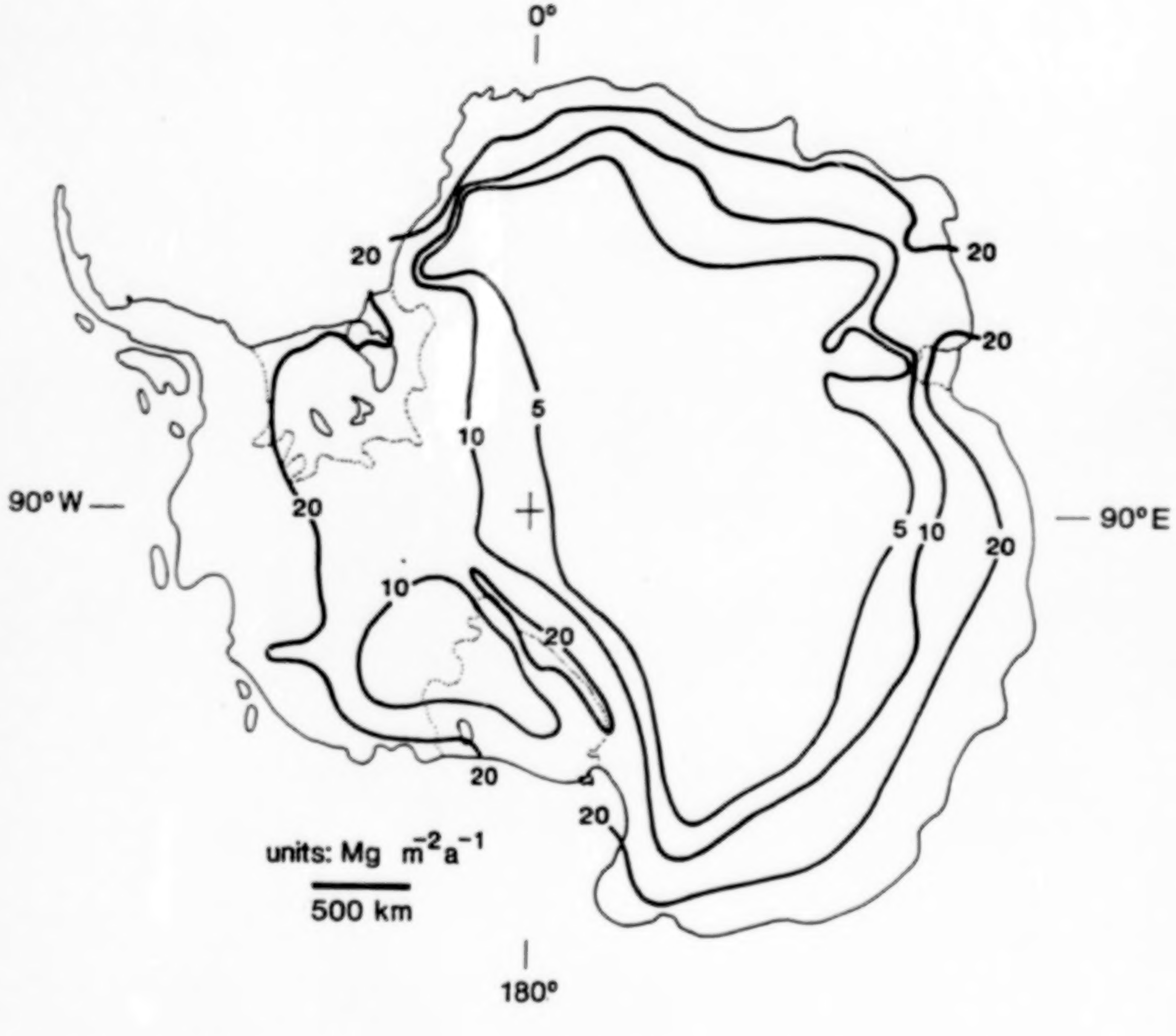


Figure 3. Net surface accumulation, from M. Giovinetto and C.R. Bentley. 1985. *Antarctic Journal of the U.S.*

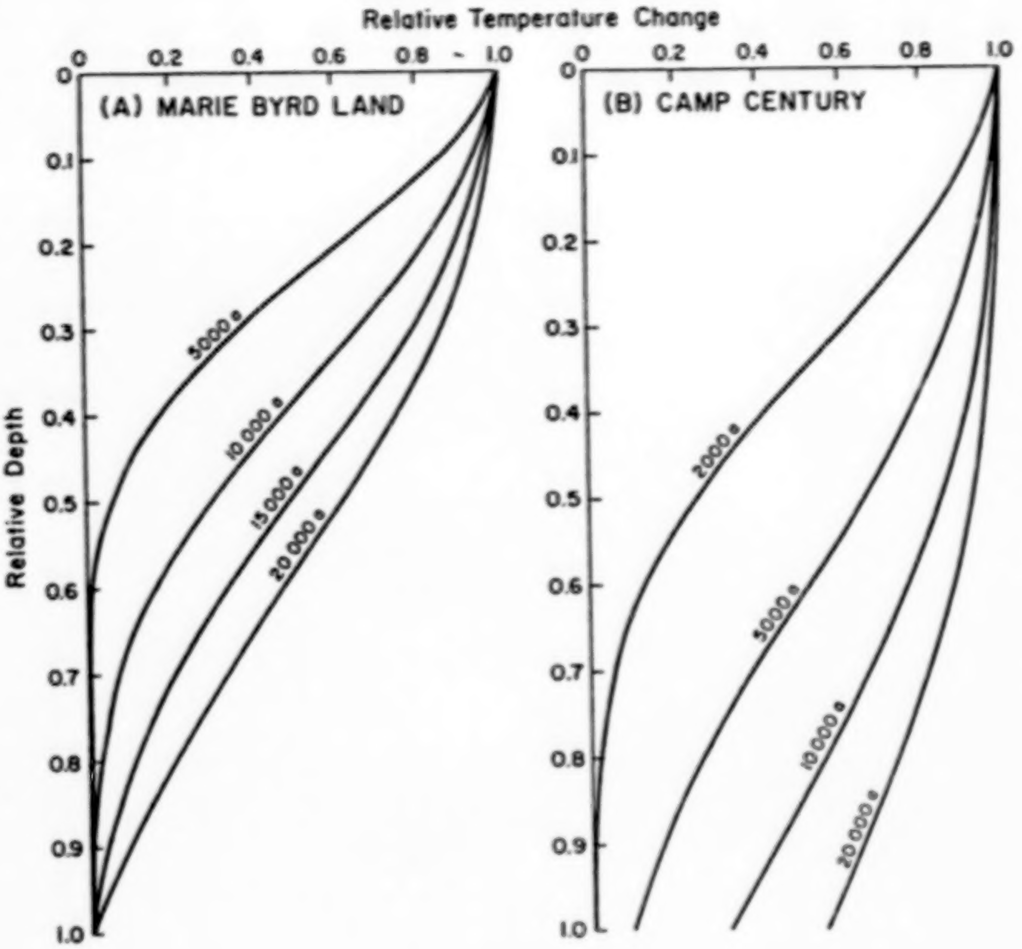


Figure 4. Propagation of surface temperature change into thick ice sheets. For Byrd Station, the temperature change at the half-depth reaches only 1/3 of the surface change after 10 000 years. The response in thin ice is much faster. Figure from Whillans, JGR, 1981.

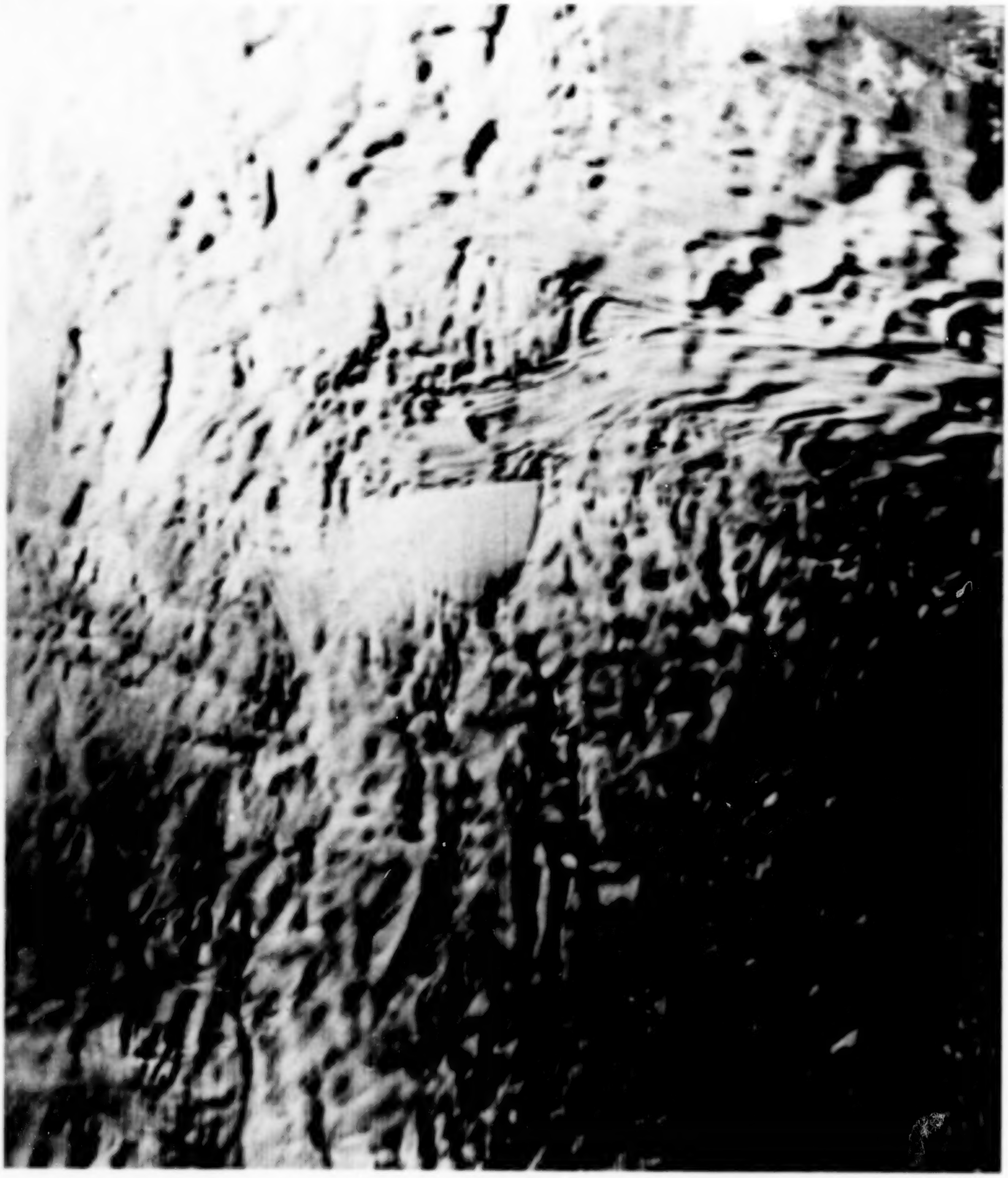


Figure 5. Landsat TM image of the onset of ice stream E. The flat half-moon feature is due to something with positive relief in the bed. It is speculatively interpreted as a table-top mountain formed by a subglacial eruption. It is about 20 km across (top-bottom). South is toward the top, and solar illumination is from the upper left. Image 014/117, personal communication from R.A. Bindschadler.

NASA
National Aeronautics and Space Administration

Report Documentation Page

1. Report No.
NASA CP-3115, Vol. 2

2. Government Accession No.

3. Recipient's Catalog No.

4. Title and Subtitle
West Antarctic Ice Sheet Initiative
Volume 2: Discipline Reviews

5. Report Date
May 1991

6. Performing Organization Code
971

7. Author(s)
Robert A. Bindschadler, Editor

8. Performing Organization Report No.
91A01040

10. Work Unit No.

9. Performing Organization Name and Address
Goddard Space Flight Center
Greenbelt, Maryland 20771

11. Contract or Grant No.

13. Type of Report and Period Covered
Conference Publication

12. Sponsoring Agency Name and Address
National Aeronautics and Space Administration
Washington, D.C. 20546-0001

14. Sponsoring Agency Code

15. Supplementary Notes
The WAIS workshop was cosponsored by NASA and the National Science Foundation, Washington, D.C.

16. Abstract

This report contains seven discipline review papers on the state of our knowledge of West Antarctica and opinions on how that knowledge must be increased to predict the future behavior of this ice sheet and to assess its potential to collapse, rapidly raising global sea level. These are the goals of the West Antarctic Ice Sheet Initiative (WAIS). A companion report (Volume 1) contains the Science and Implementation Plan of WAIS.

This initiative was discussed in an earlier report under the name of SeaRISE (NASA CP 3075).

17. Key Words (Suggested by Author(s))
Antarctica, Sea Level, Ice Sheet, Climate, SeaRISE

18. Distribution Statement
Unclassified - Unlimited

Subject Category 42

19. Security Classif. (of this report)
Unclassified

20. Security Classif. (of this page)
Unclassified

21. No. of pages
143

22. Price
A07

NASA FORM 1626 OCT 86

NASA

National Aeronautics and
Space Administration
Washington, D.C.
20546

Postage and Fees Paid
National Aeronautics and
Space Administration
NASA-451

Official Business
Penalty for Private Use $300

SPECIAL FOURTH CLASS MAIL
BOOK

L1 001 CP 3115 V2910710S002378D
GOVERNMENT PRINTING OFFICE
LIBRARY PROGRAMS SERVICE/SLIA
ACQUISITIONS SECTION
WASHINGTON DC 20401

END

3 — 23 — 92